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High-Low Inversion Point of Quartz in Metamorphic Rocks

By

Toshimichi IIYAMA

Abstract

The inversion point of quartz in metamorphic rocks ranging from the amphibolite to green schist facies and also in recrystallized xenoliths in volcanic rocks were measured. The quartz formed at higher grades in metamorphism has lower inversion points than the quartz formed at lower grades. Quartz in xenoliths has lower inversion points than quartz from metamorphic rocks and granitic rocks but has higher inversion points than the porphyritic quartz in volcanic rocks. The condition attending the recrystallization of quartz in the xenoliths is discussed.

Introduction

TUTTLE (1949) and KEITH and TUTTLE (1952), studying the relation between the inversion point of quartz and the condition of its formation, found that the quartz formed at higher temperatures has lower inversion point than the quartz formed at lower temperatures. They also pointed out that, in order to know the temperatures of formation of the rocks from such relations, it is necessary to compare the inversion point of quartz from rocks of similar chemical conpositions because the inversion temperature appears to be affected by the minor constituents contained in quartz.

TUTTLE's work was mainly on quartz from igneous rocks, and it would be interesting to know whether the inversion point of quartz varies according to the grade of metamorphism of the host rocks.

The present writer studied, after TUTTLE's method (1949), many quartz samples in metamorphic rocks of the Gosaisyo-Takanuki District, Abukuma Plateau, Hukusima Prefecture, and other localities in Japan.

The regional metamorphic rocks of the Gosaisyo-Takanuki district, were studied by MIYASHIRO (1953), who distinguished several metamorphic grades according to the natures of the minerals from the argillaceous as well as calcium-rich rocks. He furnished the writer many

samples of quartzose rocks from this area. The samples from the other localities, such as those from Tanzawa Mountainland, SW. of Tokyo, from the Sanbagawa metamorphic rocks of Titibu, NW. of Tokyo, and from the Ryoke metamorphic rocks of Akaisi Valley, Nagano and Aiti Prefectures were collected by the writer.

Method of Study

Quartz samples separated* from their host-rocks by hand picking were boiled in HCl for about 30 minutes, washed with water and dried, then they were used for measurement.

Quartz from pegmatite at Nogisawa-mura, Ishikawa-gun, Hukusima Prefecture, was used as the standard for the relative measurement of the inversion point.

The temperature was measured** by the chlomel-alumel thermocouple, made by Chinoseisakusho Co., Ltd. Whenever the electromotive forces at the inversion point of the standard quartz was found to deviate more than 5 microvolts from its original value, the couple was renewed in order to avoid the error due to the variation of its thermoelectric power. It was necessary to renew the couple in every 200 hours.

In Table 1 are given the differences between the electromotive forces (ΔT in microvolts) at the inversion point of the samples and that of the standard quartz.

Having no means of calibration of the thermocouple, the inversion temperatures are not expressed in degree centigrade. The higher positive value of the difference ΔT indicates high inversion temperature relative to that of standard quartz. Rough estimation shows that the temperature difference of 1°C corresponds to the difference in electromotive force of 40 microvolts.

The error of the inversion point measurement was probably less than 3 microvolts in ΔT.

* The grains of quartz were passed through a screen of 30 meshes and caught on another screen 150 meshes. Sometimes it was necessary to crush more finely in order to obtain clear inversion break on the time-temperature curve.

** Inversion temperature was measured only for the heating process, because it was hard to regulate the rate of cooling accurately while the rate of heating of 1°C per minute was easily maintained.

Inversion Point of Quartz and Metamorphic Grade.

In Figure 1, the values of ΔT for quartz from the Gosaisyo-Takanuki district are plotted in MIYASHIRO's geologic map showing the zonal arrangement of the rocks of the different grades. It is evident from the map that the inversion point of quartz is generally higher in the rocks of the lower metamorphic grades.

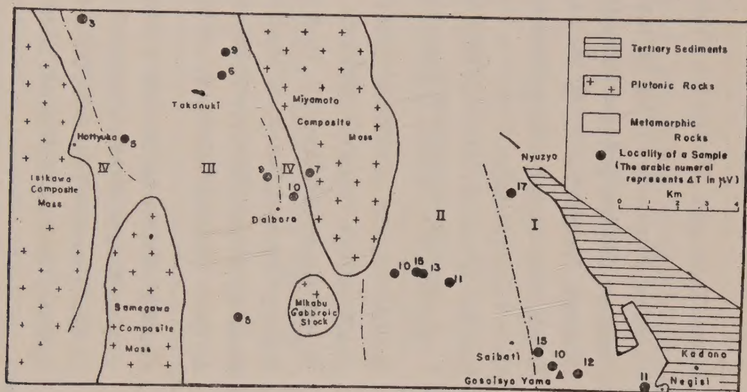

Fig. 1. Geological sketch map of Gosaisyo-Takanuki District, Abukuma Plateau, showing the inversion points of quartz in different grades of metamorphism [Geologic map after MIYASHIRO (1953).]

The Roman numerals represent the metamorphic zones.

Zone I. Characterized by the absence of clinopyroxene and grandite in calcium-rich rocks.

Zone II. Characterized by the entrance of clinopyroxene and grandite in calcium rich rocks.

Zone III. Characterized by the absence of epidote-clinozoisite as a progressive-reaction product in calcium-rich rocks.

Zone IV. Calcium-rich rocks of this zone are similar to the calcium-rich rocks of the zone III except that the rocks of this zone sometimes contain wollastonite.

In Table 1 are also shown the ΔT values for quartz from the other localities. General agreement is seen between the ΔT values in Gosaisyo-Takanuki and those of the rocks of the corresponding grades from the other localities, although the range of the ΔT values in the different grades partly overlap one another as shown in Table 2.

In Table 3, the ΔT values for quartz from some granitic rocks of Japan are given. They, ranging from -10 to $+14\,\mu V$, are roughly

Table 1. Inversion point of quartz from metamorphic rocks.

Metamorphic grade		Rock	Locality	ΔT
Green schist facies		Quartz vein cutting chlorite-graphite schist	Siozu, Kaiso-gun, Wakayama Prefecture	+20
		Quartz schist	Oyahana, Nagatoro, Titibu, Saitama Prefecture	+19
		Calcite-chlorite-quartz schist	Nagatoro, Titibu, Saitama Prefecture	+18
		Piedmontite-garnet-quartz schist	Oyahana, Nagatoro, Titibu, Saitama Prefecture	+17
		Quartz schist	Sueno, Ohno-gun, Saitama Prefecture	+16
		Chlorite-quartz schist	Near Tokusima City, Tokusima Prefecture	+14
		Quartz-calcite-chlorite schist	Kurokura, Tanzawa Mountainland, Kanagawa Prefecture	+10
		Quartz vein cutting graphite phyllite	Urakawa-mati, Sizuoka Prefecture	+10
Epidote amphidolite facies	Zone I**	Biotite-quartz schist	Nyuzyo, Gosaishy-Takanuki, Abukuma Plateau	+17
		Biotite phyllite	Gosaisyo-Toge, Gasaisyo-Takanuki, Abukuma Plateau	+12
		Garnet-quartz schist	Negisi, Gosaisyo-Takanuki, Abukuma Plateau	+11
		Garnet-quartz-biotite schist	Gosaisyo-Saragai, Gosaisyo-Takanuki, Abukuma Plateau	+10
		Garnet-quartz-biotite schist	Gosaisyo-Saibati, Gosaisyo-Takanuki, Abukuma Plateau	+13
		Garnet-biotite-muscovite schist	Nyuzyo, Gosaisyo-Takanuki, Abukuma Plateau	+12
	Zone II	Garnet-biotite schist	Gosaisyo-Takanuki, Abukuma Plateau	+15
		Biotite-muscovite-quartz schist	Isizumi-Kaiya, Gosaisyo-Takanuki, Abukuma Plateau	+13
		Biotite-muscovite-quartz schist	Kaiya, Gosaisyo-Takanuki, Abukuma Plateau	+11
		Quartz vein cutting amphibolite	Isizumi, Gosaisyo-Takanuki, Abukuma Plateau	+10
		Quartzose part in actinolite green schist	Kurokura, Tanzawa Mountainland, Kanagawa Prefecture	+11
		Quartzose part in actinolite green schist	Uenohara, Tanzawa Mountainland, Kanagawa Prefecture	+10
		Quartzite in biotite-hornfels	Near Kaso Mine, Totigi Prefecture	+10

Amphibolite facies		Sillimanite-garnet-bearing biotite-muscovite gneiss	Daibara, Gosaisyo-Takanuki, Abukuma Plateau	+10
		Sillimanite-garnet-bearing biotite gneiss	Sinokubo, Gosaisyo-Takanuki, Abukuma Plateau	+ 9
	Zone III**	Quartzite in amphibolite	Sawa, Gosaisyo-Takanuki, Abukuma Plateau	+ 9
		Pegmatitic vein through gneiss	Sakune, Gosaisyo-Takanuki, Abukuma Plateau	+ 6
		Garnet bearing quartz gneiss	Nyuzyo-Takinotaira, Abukuma Plateau	+ 5
		Quartzite	Yazi, Gosaisyo-Takanuki, Abukuma Plateau	+ 3
	Zone IV**	Quartzite	Baba, Gosaisyo-Takanuki, Abukuma Plateau	+10
		Wollastonite-diopside-garnet-quartz rock.	Daibara, Gosaisyo-Takanuki, Abukuma Plateau	+ 9
		Injection gneiss (composed of garnet, biotite, sodic plagioclase, quartz)	Sakauba, Toyone-mura, Kitasidara-gun, Aiti Prefecture	+10
		Wollastonite-calcite-quartz rock.	Hukuhara, Abukuma Plateau	+ 7
		Quartz schist in gneiss	Takato-Town, Nagano Prefecture	+ 5
		Quartzose amphibolite	Yunosawa, Tanzawa Mountainland, Kanagawa Prefecture	+ 5
		Quartz schist in gneiss	Sakauba, Toyone-mura, Kitasidara-gun, Aiti Prefecture	+ 2
		Andalusite-sillimanite-quartz vein in gneiss	Kosinmen, Gisyu-gun, Heiannando, Korea	− 4

* ΔT (in microvolts)=(E.M.F. at the inversion point of the standard sample)-(E.M.F. at the inversion point of the standard quartz)

** Metamorphic zones in Gosaisyo-Takanuki District, Abukuma Plateau (see the explanation of Fig. 1)

similar to the values of metamorphic quartz of the amphibolite facies or higher grade.

Table 2. The range of the ΔT values in the rocks of different metamorphic grades

Metamorphic grade	ΔT (in microvolts)
Amphibolite facies	+ 2 — +10
Epidote amphibolite facies	+10 — +15
Green schist facies	+10 — +20

Table 3. Some data of inversion point of quartz from granitic rocks.

Rock	Locality	ΔT
Pegmatite cutting biotite granite	Takato, Nagano Prefecture	+14
Biotite granite.	Yamagata, Kitakumi Mountainland	+13
Hornblende-biotite granite.	Takato, Nagano Prefecture	+11
Biotite granite.	Daibara, Gosaisho-Takanuki	+ 7
Biotite granite.	Urakawa, Sizuoka Prefecture	+ 6
Biotite granite.	Hukuhara, Abukuma Plateau	+ 5
Biotite granite.	Nikko, Totigi Prefecture	− 2
Biotite granite.	Sori, near Asio Mine, Totigi Prefecture	− 5
Biotite granite.	Naegi, Gihu Prefecture	−10

It may be concluded that the quartz formed at higher grades during metamorphism has lower inversion points than the quartz formed at lower grades. A relation parallel to that was found by TUTTLE (1949) for quartz from igneous rocks. The writer considers, after TUTTLE and KEITH (1953), that the overlapping in the range of ΔT values of quartz from different metamorphic grades is due to the slight difference of the minor element in quartz.

Quartzose Xenoliths in Volcanic Rocks

Quartzose xenoliths are occasionally found in volcanic rocks of Japan. They are either fragments of chert, siliceous sandstone, or granitic rocks.

Most of xenoliths of the sedimentary rocks from Norikura volcano, Nagano Prefecture and from Kumano acidic rocks, Mie and Wakayama Prefecture are made up of uniform aggregate of clear quartz grains of medium size, though some of them consist of aggregate of small

milky grains in the core and of the clear grains of medium size in the margin, the boundary between the two being gradual.

In a fragment of granitic rock of 5 cm. in diameter found in welded tuff of olivine pyroxene andesite of Nantai Volcano, Nikko, shows that a part of the quartz grain along the boundary of quartz and potash feldspar (now inverted to sanidine, though originally it was microcline.) has been melted.

Table 4. Inversion Point of Quartz from Quartzose Xenoliths in Volcanic Rocks and Quartz Phenocryst in Volcanic Rocks.

	Rock	Locality	ΔT
Xenolith	Quartzose xenolith in Augite-hypersthene andesite.	Norikura-dake, Nagano Prefecture	−13
	Quartzose xenolith in granite porphyry.	Singu City, Wakayama Prefecture	−29
	Quartzose xenolith in granite porphyry.	Takada, near Singu City	−25
	Granite fragment found from welded-tuff of olivine augite-hypersthene andesite.	Nantai Volcano, Nikko, Totigi Prefecture	−10
	(Original granite of the xenolith.	Nikko, Totigi Prefecture	− 1)
Phenocryst	Phenocryst in dacitic tuff.	Yugawara-mati, Kanagawa Prefecture	−45
	Phenocryst in granite porphyry.	Nikko, Totigi Prefecture	−30
	Average of 20 data of phenocryst in granite porphyry of Kumano acidic rocks.*		−50
	Phenocryst in biotite liparite	Kinomoto Town, Mie Prefecture	−50

* Kumano acidic rocks extend from the town of Owase, Mie Prefecture to the town of Nati, Wakayama Prefecture. The value given in the table is the average of those measured on quartz from several localities of the mass. (The actual values range from $-48\,\mu V$ to $-55\,\mu V$ in ΔT).

The ΔT values for quartz from these xenoliths are given in Table 4 together with the values for quartz occurring as phenocrysts in some volcanic rocks. The inversion points of quartz in xenoliths are generally lower than those of quartz from granitic rocks and from metamorphic rocks. This fact, taken in conjunction with the texture of the quartzose xenoliths, strongly indicates that the quartz grains have been recrystallized, when they were in wall rocks of magma chamber, or after the xenoliths were captured by the magma**. Yet, the fact that the inversion points of the xenolithic quartz are distinctly higher than those of the porphyritic quartz needs some explanation. There are two possibilities :

** This was also confirmed by Mitsue KOIZUMI who studied the decrepitation of some of the quartz samples sent to him by the writer.

(a) The recrystallization of the quartz in the xenoliths took place at some temperature below that of the surrounding magma, or at least below the temperature of crystallization of the quartz phenocrysts.

(b) The quartz in xenoliths may have recrystallized at temperatures similar to those of the formation of the porphyritic quartz. But the trace elements available to the xenolithic quartz during the recrystallization were largely from the original quartz and only partly from the magma, owing to the difficulty of diffusion. This proportion of the trace elements determined the inversion points of the xenolithic quartz.

Acknowledgements.

This study was carried out under the kind guidance of Emer. Prof. Dr. Seitaro TSUBOI of the Geological Institute, Tokyo University, to whom the writer wishes to offer his heartly thanks. Dr. Akiho MIYASHIRO of the Geological Institute, Tokyo University, placed the samples of quartz from Gosaisyo-Takanuki, Abukuma Plateau, at the writer's disposal and gave him constant advice and encouragement for the study. The writer is also indebted to Dr. Hisashi KUNO, Mr. Jiro SHIMIZU, and to Mr. Masao YAMASAKI of the Geological Institute, Tokyo University for criticism of the results and for some of the samples. The writer's gratitude is due to Mr. Mitsue KOIZUMI, Division of the Earth Sciences, Osaka University, who made the study of the decrepitation method on some of the samples. Finally, the writer's thanks are due to the Ministry of Education of Japan for the Grant in Aid for Scientific Researches which enabled this study.

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The Symplesite Problem

By

T. Ito and H. Mori†

Recently we have proposed a new mineral name, parasymplesite, for the monoclinic variety of the dimorphous minerals of the composition $Fe_3(AsO_4)_2 \cdot 8H_2O$, the other being triclinic symplesite (Ito et al, 1954). The relation that exists between the two polymorphs is analogous with that of parawollastonite and wollastonite (Ito, 1950) and the minerals provide a further example for the polysymmetric synthesis occasioned by simple gliding. The difference of them in structure arises from the difference in the amount of glide the building subunits undergo to form the unit cell.

The unit cell of parasymplesite has the dimensions (Mori & Ito, 1950): a=10.25A, b=13.48A, c=4.71A and β=108°50′. The subcell has just one half the volume of the unit cell with the b length halved and contains the whole one molecule $Fe_3(AsO_4)_2 \cdot 8H_2O$. If this subcell is subjected to an echelon gliding with a glide 2/5 a the unit cell of symplesite will ensue, while the parasymplesite cell is formed,

Table 1. Lattice constants of symplesite.

	Experimental (Wolfe)	Theoretical
a	7.85A	7.76A
b	9.39 〃	9.30 〃
c	4.71 〃	4.71 〃
α	99°55′	101°25′
β	97°22.5′	95°44′
γ	105°57.5′	104° 6′

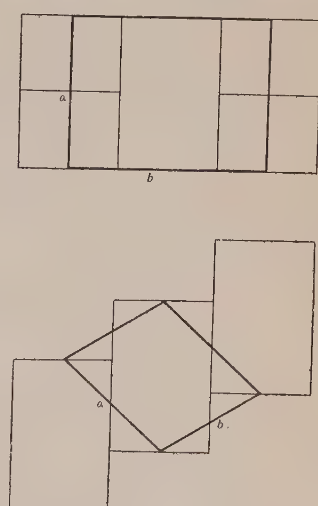

Fig. 1. The relation of parasymplesite and symplesite illustrated. The same subcells (*thin lines*) produce by a glide $a/2$ the parawollastonite cell (*upper figure, thick lines*) and by a glide $2a/5$ the symplesite cell (*lower figure, thick lines*) (Projection parallel to the c axis on (001)).

† Deceased 21 June 1954.

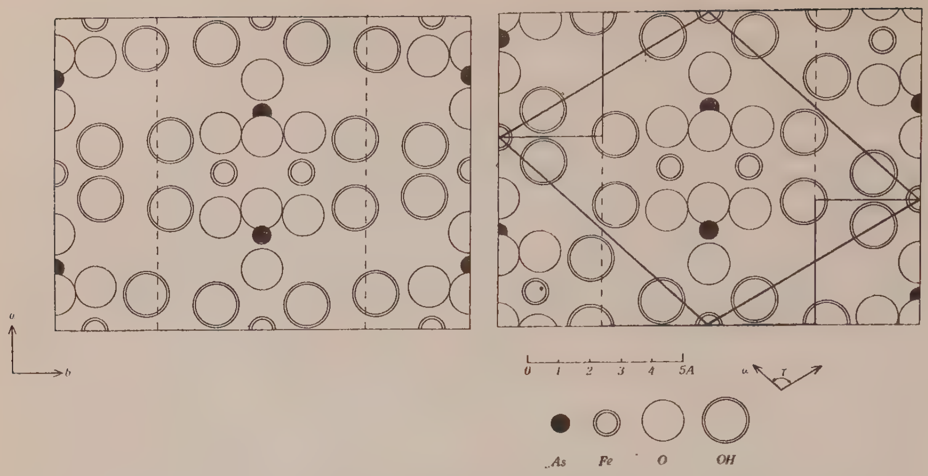

Fig. 2. The structure of parasymplesite (*left*) and the ideal structure of symplesite (*right*) derived from it in the way as explained in the text and illustrated in Fig. 1. (Projection parallel to the *c* axis on (001)).

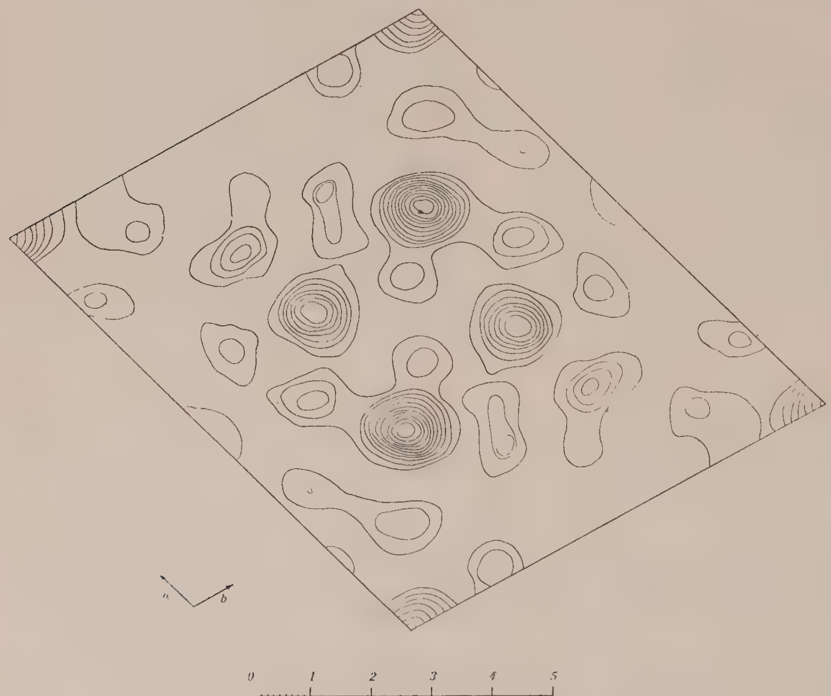

Fig. 3. Fourier projection of electron density of symplesite on (001), with contours at intervals of 5 e.A^{-2}, the zero-electron lines being omitted.

as is always the case with a C-centred lattice, by either an echelon or alternate gliding with a glide $a/2$ (Fig. 1). We give in Table 1 the lattice constants of symplesite obtained experimentally by WOLFE (1940) compared with those derived from parasymplesite in this way. The matrix of transformation from the monoclinic to triclinic axes is: $\frac{3}{8}\frac{1}{2}0\Big/\frac{5}{8}\frac{\bar{1}}{2}0\Big/001$.

The structure of parasymplesite (worked out formerly as of symplesite by MORI & ITO (1950)) is built up of complex bands of the composition $Fe_3(AsO_4)_2 \cdot 8H_2O$. These bands are joined to each other by the weak bonds due to the tetrahedral arrangement of H_2O molecules that border the bands. In the derived structure of symplesite this mode of linkage is for the most part maintained. The only change is that the grouping of H_2O molecules is localized at the different points of band peripheries. Moreover, the H_2O tetrahedra

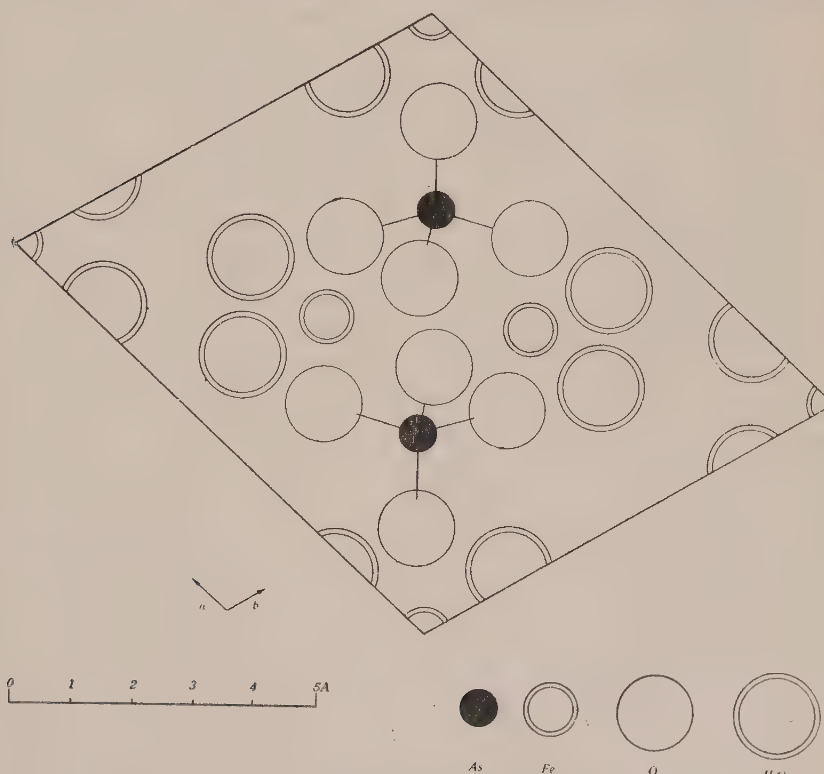

Fig. 4. The structure of symplesite projected on (001).

are not so regular as in parasymplesite. A consequence of the (unsymmetrical) echelon gliding in symplesite is the loss of symmetry planes (Fig. 2).

In the actual structure of symplesite as determined by a two-dimensional Fourier synthesis on (001) further shifting of the positions of polyhedra of oxygen and oxygen-OH around As and Fe atoms takes place (Fig. 3). Fig. 4 illustrates the structure of symplesite as projected on (001). The data for the structure will be published elsewhere later.

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On the Tectonic History of Taiwan (Formosa)

By

Teiichi KOBAYASHI

In a series of the Ryukyu and northern festoon arcs in the north-western side of the Pacific basin virgation is undeveloped, while it is typically represented in another series comprising the Philippines and southern islands. Thus the two series indicate a remarkable contrast in the geologic structure of Eastern Asia. As noted in my "*Sakawa Cycle*", 1941, Taiwan or Formosa, 25,570 square kilometers, attracts special attention as their link. As I had a happy opportunity to see the general picture of her geology in my return trip from the VIII Pacific Science Congress at Quezon City, Philippines, 1953, I take the pleasure of discussing the tectonic development of the island. On this occasion I wish to express my most cordial thanks to Prof. Ting Ying H. MA of the National University of Taiwan, Messrs. L. S. CHANG (formerly, Reikyoku CHOH), T. P. YEN (Soha GAN) and T. L. HSU of the geological survey of Taiwan for their guidance in my excursions and to the survey and other branches of the Chinese government and mining and other companies for the facilities of my visit.

Although some geological observations were reported since the middle of last century, they were very fragmentary at the beginning and Taiwan has long remained as a *terra incongnita*. Her geology was, however, immensely clarified in a half century from 1895 to 1945. During this epoch the geological sheet-maps in the northern part (1/50,000 and in part 1/1,000,000) and the geological survey of the Neogene oil fields were accomplished. Beside them many valuable contributions were done to her geology, both on the basic and applied sides. Among the recent advancements more important are the discoveries of Permian and Upper Cretaceous fossils in the metamorphosed complex by YEN (1951) and of the Oligocene fauna by CHANG(1953).

The synthesis attempted here is of course based on a mass of data accumulated through these years. Their interpretation, and coordination in the historical development and the tectonic lineament of the island in geology of Eastern Asia is made by my own judgement. In my

opinion the metamorphosed Permian and older rocks belong to the axial zone of the Cretaceous Sakawa folded mountains. The Taiwan subgeosyncline was brought forth on the continental side of this axis. The Triassic Akiyoshi folded mountains further beyond were its foreland. (See Fig. 1).

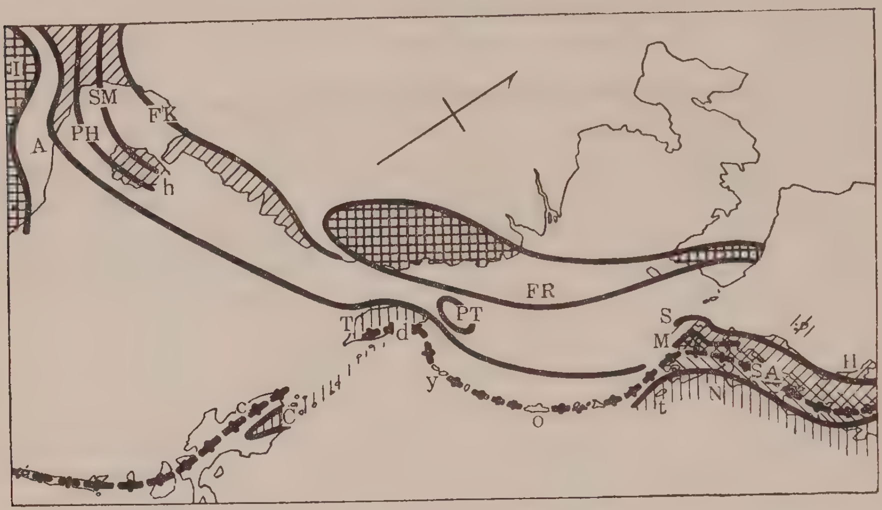


Fig. 1. Tectonic Position of Taiwan in Eastern Asia.
FR. Fukien-Reinan massif. FK. Fansipan-Kuantung metamorphosed axis.
SM. Song Ma arc with R. Noire nappe. PH. P'ou Huat arc.
A. Annam zone. I. Indosinian massif. PT. Pre-Taiwan dome.
H. Hida gneiss zone of Akiyoshi folded mountains
S. Sangun principal metamorphic zone of Akiyoshi folded mountains.
M. Motoyama auxilliary metamorphic zone of Akiyoshi folded mountains
SA. Axis of Sakawa folded mountains. T. Taiwan folded zone.
C. Cagayan synclinorium. N. Nakamura folded mountains.
t. Tanegashima. o. Okinawa-jima. y. Yaeyama insular group.
d. Dainano metamorphic zone. c. Cordillera Central.
h. Hainan island (海南島).

Neither angular discordance nor orogenic sediment is found in the Tertiary sequence except the middle Neogene Koshun conglomerate in southernmost Taiwan. Thus I failed to convince myself of the Oligocene or middle Tertiary orogeny or related metamorphism as emphasized by many geologists among which HAYASAKA and others (1947) and YUAN (1953) are the latest. It is my opinion that the middle Tertiary event was the westerly shifting of the sinking axis, caused by upheaval of the embryonic anticline in the axial zone. Because there is no sharp boundary between the metamorphosed Palaeogene and non-

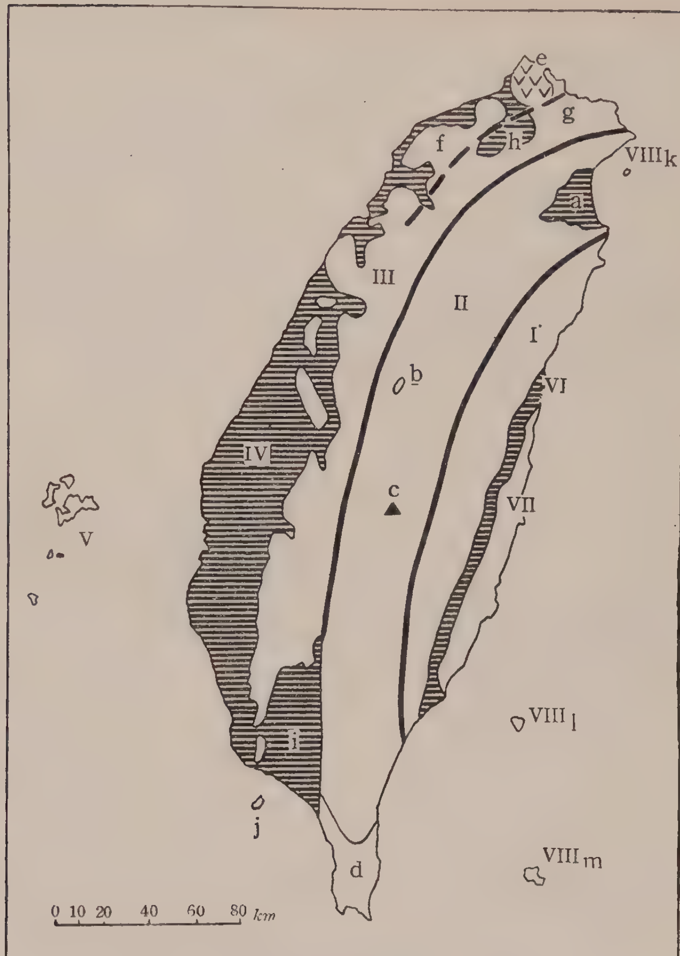


Fig. 2. Teconotic division of Taiwan.

I. Dainano zone
II. Urai zone
  a. Ilan (Giran, 宜蘭) plain
  b. Puli (Hori, 埔里) basin
  c. Mt. Niitaka (新高山)
  d. Southernmost Neogene
III. West Taiwan zone
  e. Daiton volcanic group
  f. Toen undulated subzone
  g. Kiirun imbricated subzone
  h. Taipei (Taihoku) basin
  i. Pingtung (Heito, 屏東) plain
  j. Liukiuyü (Ryukyusho, 琉球嶼)
IV. Western coastal plain
V. Penghutao (Boko island, 澎湖島)
VI. Kwarenkei rift valley
VII. Daito coastal range
VIII. Eastern dependent islands
  k. Kueishantao (Kizanto, 亀山島)
  l. Huoshaotao (Kwashoto, 火焼島)
  m. Huangt'ouyü (Kotosho, 紅頭嶼)

metamorphosed Neogene, the metamorphism commenced probably in the middle Tertiary, is thought to have continued by the end of the Plio-Pleistocene paroxysm of the *Taiwan orogeny*.

The gentle foldings or warpings were repeated in the geosyncline. The mountain structure constructed at length by the orogeny consists of the Dainano, Urai and West Taiwan zones in addition to the Kwarenkei rift valley and Daito coastal range. The former three zones indicate the western wing of the Taiwan anticlinorium which are separated from one another by tectonic lines. On the two sides of the Daito coastal range there are still greater rupture lines with which its eastern wing was destroyed. (See Fig. 2).

Like the south bent of the Japanese arc in Kyushu the Ryukyu arc is abruptly bent to the south in North Taiwan. At this back bending the Taiwan geosyncline was strongly compressed with the result that thrustings were repeated against the foreland and a typical imbrication was introduced. There the Kiirun imbricated subzone and the Toen undulated subzone can clearly be distinguished in the west Taiwan zone. Subsequently, however, Taiwan, was detached from the Ryukyu islands and performed geanticlinal upheaval by itself, causing dislocations among the tectonic zones. Because the geanticline was asymmetrical, the backbone range runs close to the eastern coast, Mt. Niitaka, 3,950 m. high in the Urai zone being the highest.

Penghutao isles to the west of Taiwan are ruins of a Pleistocene basaltic mesa on the Akiyoshi terrain. The eastern chain of dependent isles lie on the same submarine ridge with Batan, Babuyan and other isles of the Philippines which is a northern projectile from the Tertiary synclinorium of the Cagayan valley in Luzon and is separated from the Daito coastal range by a deep submarine trench.

Lately I have demonstrated that the Triassic Akiyoshi folded mountains on the inner side of Japan extend into Tonkin, Indochina through Hainan island (1951). Prior to this I have concluded that the Cretaceous Sakawa folded mountains in the median part of Japan is traceable into the axial zone of the Ryukyu islands (1941). Permian fusulinids were discovered by HANZAWA (1935) in a limestone in the northwestern peninsula of Okinawa. In the main part of the island the Palaeozoic formation is represented mainly by shale and sandstone, and partly metamorphosed into chlorite schist and graphite schist. In addition, however, there are limestone lenses, Radiolarian chert layers and conglomerate beds containing exotic granitic rocks like the Permian

Usuginu conglomerate in Japan. The Palaeozoic formation is intruded by diabase or granite in Tokunoshima and Ishigaki-jima. In Ishigaki and Kohama isles in the Yaeyama group there are phyllites and crystalline schists which indicate the axial core of the Sakawa folded mountains.

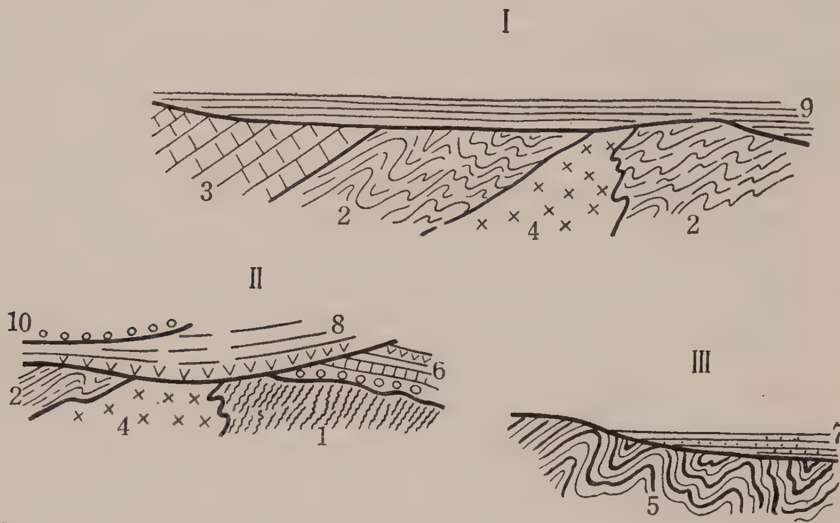


Fig. 3. Idealized sections of the Ryukyu arc showing the stratigraphic relation.
I. Okinawa-Oshima group.
II. Yaeyama group.
III. Tanegashima.

10. Plio-Pleistocene Sonai conglomerate 9. Pliocene Shimajiri formation
8. Miocene Yaeyama formation 7. Lower Miocene Kukinaga formation
6. Eocene Miyara formation 5. Palaeogene and older Kumage formation
4. Late Mesozoic (?) intrusive rocks 3. Permian limestone
2. Upper Palaeozoic formation 1. Crystalline schists

In Tanegashima in the north there is the strongly folded Kumage shale and sandstone formation, Palaeogene and older in age, on which lies the Kukinaga sandstone and conglomerate formation containing *Vicarya callosa*. Therefore the Kumage, combined with the folded pre-Miocene Nakamura group in South Kyushu, Shikoku and the further east represents the Tertiary Nakamura folded mountains.

Merit to the discovery of Permian fossils it became evident that the metamorphosed Dainano formation indicates the extension of the Ryukyu axis, i. e. the Sakawa axis. In the Dainano metamorphosed zone there are three distinct belts, namely (1) belt of sericite chlorite

schist and chlorite schist, (2) thick crystalline limestone belt and (3) belt of chlorite schist and graphite schist enumerated from the east. In the third belt there are, in addition, quartz schist, quartzite and limestone lenses in the last of which YEN discovered Permian fusulinids and tetracorals which were respectively determined by THOMPSON and MA. These metamorphics are intruded by diabase and gabbro at some places and partly altered into gneiss by injection of granitic magma. Like the Nagatoro metamorphic group in the Sakawa axis of Japan, *Kieslager* of Besshi type is imbedded in the Dainano group.

Stratigraphicalcal Sequence of Taiwan

Alluvium	Rinko alluvium (Taipei, or Taihoku basin)	
	Lower terraces	
Diluvium	Upper terraces	
	Tableland gravels, −30 m.	
	Ryukyu limestone	
	Tokasan series, ca 1500 m.	Shokkozan conglomerate Kozan sand
Pliocene	Byoritsu series	Takuran beds, ca, 1400 m. Kinsui shale, +500 m.
Miocene	Kaizan series	Keichikurin beds, +500 m. Upper coal measures, −500 m. Upper shell beds. −800 m. Middle coal measures, +450 m. Lower shell beds, ca. 500 m. (Kokan tuff) Lower coal measures, −500 m. (Seitan beds)
Oligocene	Urai series	(Lowest coal measures) Suichoryu beds, +2000 m. Yonryo sandstone, ca. 500 m. } Suo series Nishimura beds, +500 m.
Eocene		
Up. Cretaceous		Pihou series
Palaeozoic		Dainano series

The schists and granitic rocks are discordantly overlain by the basal conglomerate of the Pihou formation. Its Upper Cretaceous age is determined by MA's study on hexacorals which were discovered by YEN in

a thin limestone lens. This formation is overlain unconformably by the Suo formation with a conglomerate at its bass,which is composed mostly of black phyllitic clayslate and quartzose sandstone, and its calcareous layers above it yield Eocene foraminifers. Because the above mentioned sequence seen in the northern part of the Dainano zone is reverse, they must be located in the lower wing of an overturned anticline near the western margin of the zone. Little is known of the rest of the metamorphosed terrain because travelling has been dangerous.

While the Neogene is extensive in the west Taiwan zone, the Urai zone is mostly occupied by the Palaeogene Urai formation. Clayslate and shale are leading components of the latter which can be divided into three parts with the Yonryo sandstone in the middle. *Camerina* and other Eocene fossils have long been known to occur at some places. According to CHANG's recent study the Yuhangian foraminifers from the upper or Suichoryu division is, however, Oligocene.

In North Taiwan four coal measures are intercalated in the formations from Upper Oligocene to Miocene. They are inserted within neritic sediments, suggesting four minor cycles of sedimentation due to pulses of embryonic folding. There were some volcanic eruptions within the Tertiary geosyncline which were fairly strong in North Taiwan in the Miocene, as indicated by basalt flows, tuffs and agglomerates of Kokan on the lower coal measures. (See Fig. 4).

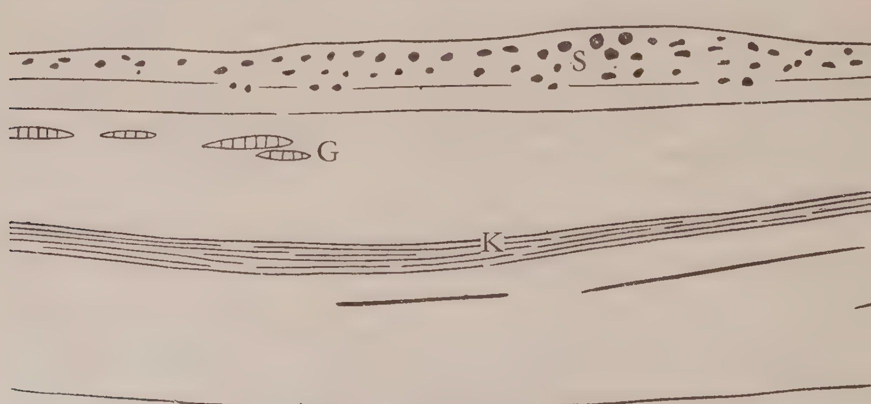
The lowest coal measures in the Suichoryu division are restricted to occur in the most northeastern part, while the lower coal measures are more extensive in North Taiwan. The middle coal measures are traceable as far south as Sinchu (新竹) and Chunan (竹南) and only the upper ones are distributed farther south to Alishan (阿里山). Reciprocally to such a southern advance of coal measures, limestone lenses show the tendency to retreat. More precisely, Miocene *Lepidocyclina-Miogypsina* limestone occurs in North Taiwan (Taipei and Sinchu) and Pliocene *Gypsina* limestone in Central and South Taiwan (Tainan and Kaohsiung, 高雄).

Compared to the Palaeogene, alternations between shale and sandstone are remarkably developed in the Neogene formation. Because staurolite, garnet, purple zircon, monazite and several other heavy minerals in the sediments are exotic for Taiwan, Fukien massif was suggested by ICHIMURA for their provenance. It may be, however, better elucidated, if a *pre-Taiwan dome* is assumed in the foreland or in

the Akiyoshi axial zone which must have been exposed before the sediments of the Hwangho, Yangtze and other rivers levelled the basement relief of the China sea. This is because high grade minerals like staurolite are known to exist in the Hida gneiss group in Japan which belongs to the Akiyoshi metamorphics. Furthermore, in this kind of folded mountains it is recognized as a tendency for the foreland to warp up by embryonic folding, for example, the Ozark dome in front of the Ouachita mountains, and the Nashville, Cincinnati and Adirondack domes in front of the three arcs of the Appalachian mountains. The supposed dome must have been located to the north of Taiwan because a similar mineral assemblage is found also in the Miocene coal measures in the Yaeyama group.

Glauconitic sandstone is distributed in various Tertiary horizons and becomes more common upward, but totally absent in the Suichoryu and older formations. It is an authigenic product under agitation of a shallow sea. Together with foraminiferous or coralline limestones and false-bedded sandstone it suggests a shallow sea for the Neogene geosyncline. The Kinsui shale extensive in the lower part of the Pliocene merges up with the Takuran alternation in which sandstone exceeds shale. It is thicker in the south near Tainan where the thickness of the alternation measures 2,500 m. The distribution of *Ostrea* banks appears to show the retreat of the sea toward the south.

Thus there is no orogenic sediment in the above mentioned Tertiary sequence, although some conglomeratic sandstone containing small pebbles are found in the Neogene formation. As discussed later, the Koshun conglomerate in the southernmost part of the Urai zone



is a single exception. No clear-cut clino-unconformity has not as yet been found in the Tertiary sequence of Taiwan.

The Plio-Pleistocene Tokasan formation which is dated by the mammalian fauna, is a typical orogenic sediment. It is generally conformable with the Pliocene, though local or minor erosion-unconformity is seen in rare instances. It consists of Kozan sandstone facies and Shokkozan conglomerate facies where the former is on a whole higher and the latter lower, but the two easily merge or interfinger with each other laterally.

The Kozan facies is chiefly composed of fine muddy silt or sand beside intercalations of clay and gravels; limonite often contained and false-bedding common. A copious fauna comprising echinoids and molluscans is known from North Taiwan.

Though sandstone wedges and lenses are not uncommon in the other facies, conglomerate is its principal member. Hence the name, Shokkozan conglomerate. Drift woods are occasionally met with and *Unio*, *Corbicula* and other shells contained therein. The gravels attain over 1 meter in diameter at the maximum, but most of them are in size of fist to man's head and well rounded. Its thickness exceeds 1,000 meters near Taichun in Central Taiwan, but only 20 m. at the minimum in North Taiwan.

It is reasonable to consider that the Shokkozan conglomerate is the sediments of deltas and fans on the west side of the backbone range at the time of its upheaval, because it is thick in Central Taiwan where the range is highest and because its material is mostly derived from the Urai zone. Because the Tokazan formation is folded

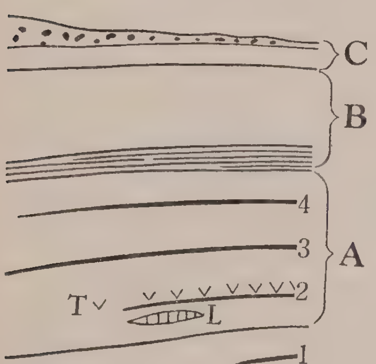


Fig. 4. Diagramatic meridional section showing the lateral change of facies and thickness of the geosynclinal sediments in the west Taiwan zone from the north (right) to the south (left).

C. Plio-Pleistocene
B. Pliocene
A. Miocene
S. Shokkozan conglemerate facies
K. Kinsui shale
T. Kokan tuff.
G. *Gypsina* limestone
L. *Lepidcoyclina* limestone
4. Upper coal measures
3. Middle coal measures
2. Lower coal measures
1. Lowest coal measures.

to some extent, it is certain that the crustal disturbance did not cease by the end of the Shokkozan age.

As shown in a series of profiles through North Taiwan, the geologic structure from the Dainano to the west Taiwan zone is a grand anticlinorium with the Sakawa metamorphics at its core. The metamorphism becomes lessened distally. (See Figs. 5 and 6).

Much cannot as yet be mentioned of the structure of the Dainano zone, but it is known that an overturned anticline near the western margin thrusts itself upon the Urai zone along the Sanseizan tectonic line. In the Urai zone about 30 km. in breadth, foldings and thrustings are repeated, the folded axes or thrust planes being inclined to the southeast in different degrees. The most important key bed is the Yonryo sandstone which shows the general tendency for the folds to

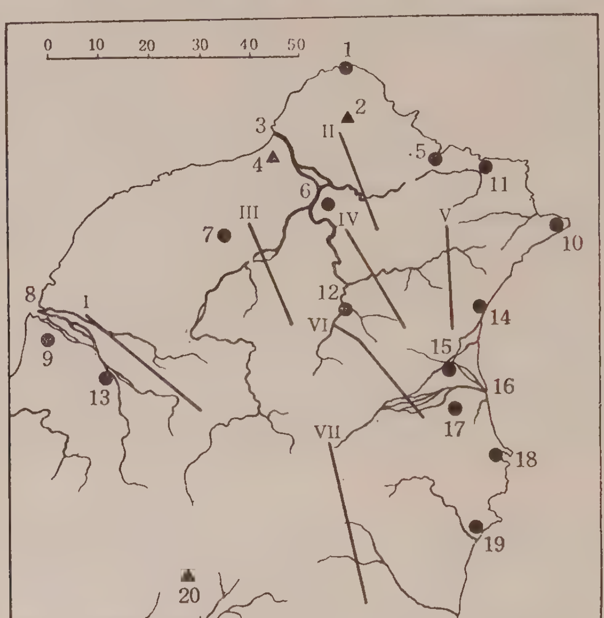


Fig. 5. Index map of the profies through North Taiwan.

1. Fukueichio, Fukikaku, 富貴角 2. Tatungshan, Daitonsan, 大屯山
3. Tanshuiho, Tansuigawa, 淡水河 4. Kuanyinshan, Kannonsan, 観音山
5. Keelung, Kiirun, 基隆 6. Taipei, Taihoku, 台北
7. Taoyuan, Toen, 桃園 8. Fengshanchi, Hozankei, 鳳山溪
9. Hsinc'hu, Shinchiku, 新竹 10. Santiaochio, Sanshokaku, 三貂角
11. Juifang, Zuiho, 瑞芳 12. Wulai, Urai, 烏来 13. Chutung, Chikuto, 竹東
14. Touwei, Toi, 頭囲 15. Ilan, Giran, 宜蘭 16. Choshuichi, Dakusuikei, 濁水溪
17. Lutang, Lato, 羅東 18. Suao, Suo, 蘇澳 19. Tananao, Dainano, 大南澳
20. Mt. Tsugitaka, 次高山.

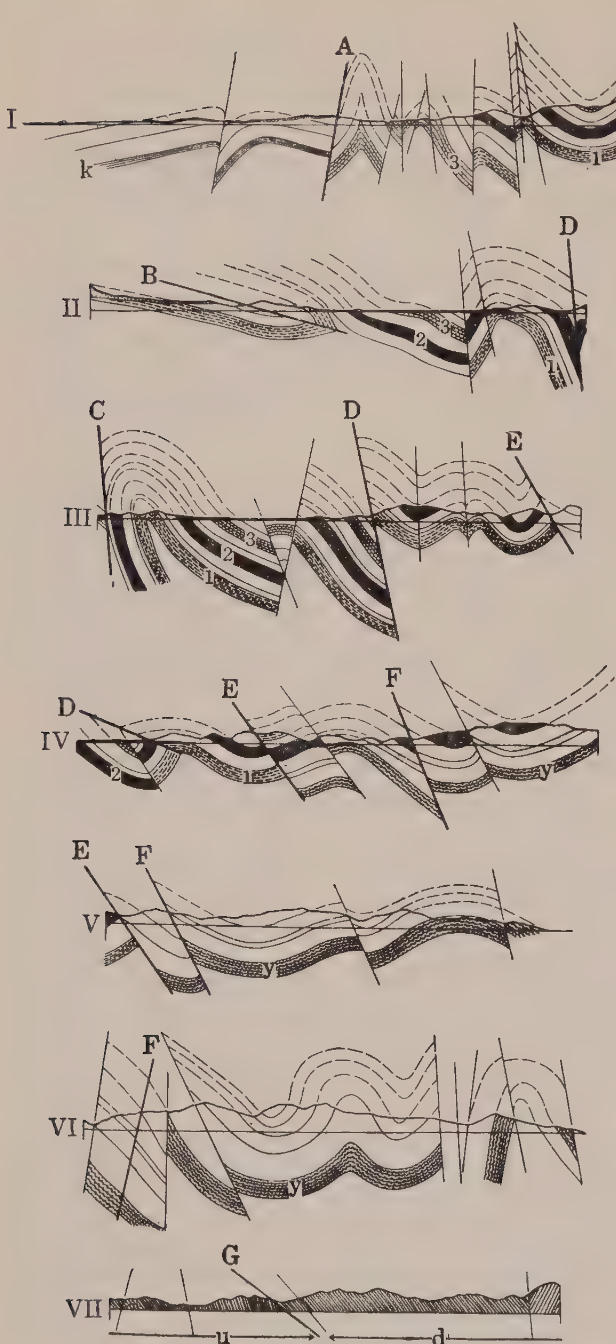


Fig. 6. Profiles through North Taiwan.

I–V. From the sheet maps of Chikuto, Taihoku, Toen, Shinten and Toi by Y. Ichikawa (1930–34).

VI. From Giran sheet map by M. Usami 1936.

VII. From Dainano sheet map by M. Ogasawara, 1933; Scale different from the precedings.

A. Chikuto (竹東) fault.
B. Kankyaku (崁脚) fault.
C. Shinso (新莊) fault.
D. Shinten (新店) fault.
E. Kushaku (屈尺) fault.
F. Kinkwaryo (金瓜寮) fault.
G. Sanseizan (三星山) fault.
k. Kinsui shale
3. Upper coal measures
2. Middle coal measures
1. Lower coal measures
y. Yonryo sandstone
u in VII. Urai zone
d in VII. Dainano zone

be more compressed on the eastern side where the folded axis or thrust plane is subvertical. On the west side on the contrary the folding is more gentle, but the thrusts are more low angled.

The Kushaku thrust draws the boundary between the Urai and west Taiwan zones, but its amount of displacement is not much different from the others in the Urai zone. In the Kiirun subzone, however, the Miocene is extensive and strongly compressed in form of imbrication; in the Toen subzone on the other hand the Pliocene extensive and gently folded. Their boundary is sharply demarcated by a low angle thrust on the north side of the Taipei basin and a few Klippen are found in its northeastern extension. The gentle undulation of the Toen subzone suggests its being the marginal part of the foreland.

Near Chutung (竹東), however, the boundary fault is normal and its downthrow on the northwest side. Because the embryonic folding in the Kiirun subzone was composed of brachyanticlines and brachysynclines arranged somewhat in checker pattern, their end-products are arcuate thrusts in a similar pattern. Their planes inclined in different degrees between the median and lateral parts of each arc; breadth of the thrusting sheets changeable. The imbrication is cut by strike faults as well as diagonal or rectangular ones. The latter is mostly reverse but the former normal. These fault systems suggest that the folding and thrusting of the geosynclinal sediments gradually developed into a compressive block movement of the already imbricated terrain in which the basement blocks probably played a role. Later, however, there took place a tension movement and the triangular Taipei basin was produced.

The two subzones are distinctly separated by a boundary thrust in North Taiwan where the west Taiwan zone is sharply bent, but in Central and South Taiwan the two subzones are inseparable because the imbricated subzone merges with the zone of open folds with subvertical axes.

The Chuhuangkeng (出磺坑) oil field is particularly interesting for analysis of the tectonic development. There is a long anticline which is, however, bisected by an axial fault along which the eastern wing is a little sunk down. As it is a tension fault at the saddle, the displacement is presumably reduced downward. In the eastern wing, however, there is a thrust of greater displacement, and the

same *Operculina* sandstone occurs twice. It is probably a dislocation in the course of the folding. The axial fault on the contrary may be the final adjustment. (See Fig. 7).

The southern end of the Chuhuangkeng anticline is cut by a transverse thrust called Sansa (三叉) along which the southern block thrusts itself upon the other. This thrust is cut by Shinkai (新開) and Dora (銅羅) reverse faults at its lateral ends. Thus the folding developed into a compressive block movement which yielded the Sansa thrust and later the two others came out. Along the Dora fault the dislocation appears to have recurred after the deposition of the tableland gravels.

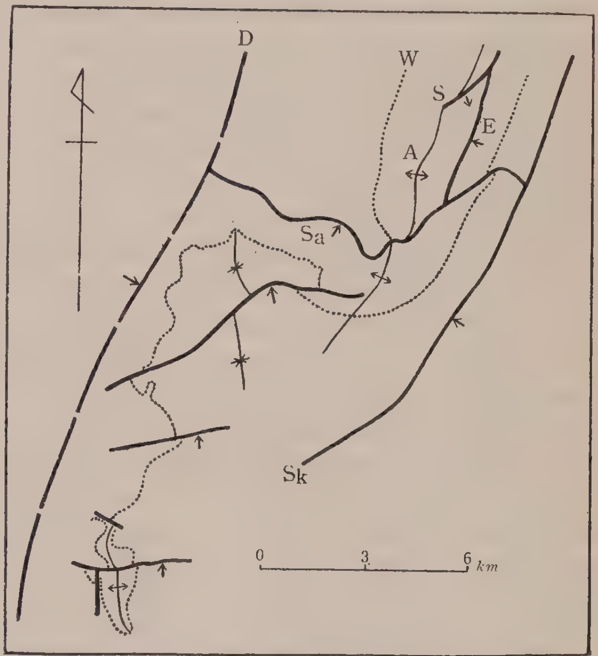


Fig. 7. Tectonic map of the Chuhuangkeng (Shukkoko) oil field (after CHANG).
W. White sandstone A. Shukkoko anticline
S. Shukkoko axial fault E. East Shukkoko thrust
Sa. Sansa thrust Sk. Shinkai reverse fault
D. Dora reverse fault.

Seeing the differential movement among the brachyanticlines and brachysynclines and in considering their development from embryonic domes and basins aligned in checker pattern, it is quite probable that the deformation of the geosynclinal sediments was controlled by the frame work of their basement which belongs to the inner side of the Sakawa folded mountains. It is well known in Japan that her inner zone was already destructed into a faulted mozaic by block movements in the later Cretaceous and early Palaeogene periods.

Because the repetition of the embryonic foldings is well shown by the middle Tertiary minor cycles of sedimentation in the Kiirun imbricated subzone and because neither the Shokkozan conglomerate

nor the Pliocene is undeveloped there, the imbrication has been completed possibly before the main upheaval phase of the backbone range.

After the fundamental anticlinorium of Taiwan had been built up with the Sakawa metamorphics at its core, the Urai zone, instead of the Dainano zone, was most elevated. The upheaval in this phase attained at the maximum in Central Taiwan, instead of North Taiwan. The tectonic analysis in the west Taiwan zone has shown that the folding was followed by the compressive block movement. On the basis of these facts it is reasonable to consider that, caused by the geanticlinal upheaval, differential movement has taken place among blocks. The dislocation must have been especially great at the boundaries among the three tectonic zones of different plasticity. The displacement along the western boundary of the Urai zone was greater in Central than in North Taiwan, as can be recognized by a comparison between the Kushaku thrust in North Taiwan and the Nansei fault in Central Taiwan. It is still a question whether the boundary is marked off by a continuous tectonic line or whether it is represented by a series of faults and thrusts en échelon. At all events it is certain that the orography was greatly changed by the upheaval of the backbone range.

As the result of such a great upheaval the Urai zone was strongly eroded. The clastic Urai rocks in a huge amount were transported beyond the western scarps through steep valleys. The Shokkozan conglomerate is such an orogenic sediment of tremendous thickness which was accumulated in a short time on the western lowland.

Palaeo-Taiwan at this time was a peninsula connected with the continent by a land bridge probably through North Taiwan and pre-Taiwan dome. The Plio-Pleistocene mammals migrated through this bridge into the western lowland where the Shokkozan fans and deltas were alinged. The bay further to the west was extended as far as the neck of the peninsula.

On the east side of the backbone range there is the Daito coastal range which is composed of Neogene formations containing a greater amount of volcanic and pyroclastic materials than those of the west Taiwan zone. The formations are moderately folded and cut by faults. The base of the Neogene is unexposed. The coastal range is detached from the Dainano zone by the Kwarenkei rift valley. Whether it is a graben or not cannot actucally be seen because the rupture lines lie concealed beneath young sediments.

Off the east coast there are three groups of isles which are mostly

composed of volcanic and pyroclastic materials. Quartz-schist caught in andesite as a xenolith in Kueishan (亀山) isle is a proof of linking of the Sakawa mountains from the Dainano zone to the Yaeyama group. Huoshaotao (火焼島) and Huangt'ouyu (紅頭嶼) are remnants of submarine volcanoes and in the latter Aquitanian limestone is inserted in andesitic tuff-agglomerate.

As noted already, these isles form a volcanic chain with the northern dependent isles of Luzon. They as a whole lie on the submarine projectile from the Cagayan valley. The Neogene formations in addition to the upper Oligocene Ibulao limestone in the valley form a synclinorium between the Sierra Madre range on the east and the Cordillera Central on the west side. Like the Yaeyama coal-bearing, the lower Miocene Lubugan is a coal-bearing formation. The middle Miocene Tugaegaro sandstone becomes tuffaceous in the south till at length it merges with volcanic rocks.

In the Philippine islands the middle Miocene, frequently quite thick, must be a sediment of the orogenic epoch. Conglomerate is well developed. Clino-unconformity is extensively seen between the middle and upper Miocene formations, although the latter is mostly composed of fine clastic rocks and limestones and the occurrences of conglomerate is restricted.

Because the Cordillera Central belongs possibly to the Sakawa axis and because the middle Miocene attains such an extraordinal thickness as 3,000 m. in the Bueda valley near the southern end of the range, the Koshun conglomerate must be discussed again.

In the southernmost part of the Urai zone TAN discovered a phyllitic conglomerate containing clayslate and another containing *Discocyclina* limestone. These boulders are probably the upper Urai members, showing intermittent emergences.

The Urai formation is in fault contact with the Koshun formation which is very thick and consists mostly of sandstone and conglomerate in alternation. In its lower part, however, there is shaly facies in which *Operculina ammonoides* was found. Therefore the formation may be upper Miocene and younger. The provenance of exotic alkaline granite and some other igneous rocks in the Koshun conglomerate is still unknown, but it can be said that the emergence indicated by the conglomerate is not much different from that of Cordillera Central in age. The adjacent land was undoubtedly high in the middle Neogene. In

the vicinity of Mt. Niitaka, however, the Urai formation is disconformably overlain by Miocene sandy shale containing *Operculina venosa*.

If these conglomerates are excluded, the Tertiary of Taiwan is composed of fine rocks in main and the appearance of the Shokkozan conglomerate is quite sporadical. The next younger is the Ryukyu limestone which is the oldest among the raised coral reefs and distinctly tilted, while the younger reef limestones in three series are all subhorizontal. At Kotobuki hill (寿山) of Kaoshiung, 356 m. high. in South Taiwan, the Takuran formation is capped by the Ryukyu limestone clino-unconformably.

In West Taiwan hills, 300 m. or less above the sea, are extensively covered by gravel beds about 30 m. thick. Because their top is flat, they are called "tableland gravels", but it happens also that the flat top is slightly tilted or gently undulated. Gravel beds are distributed also in some intermontane basins. In the Puli (埔里) basin the gravel beds attain as high as 700 m. above the sea. Beneath them there are lignite bearing lacustrine clay beds which are gently inclined and may be approximate to the Ryukyu limestone in age.

It is an interpretation that the gravel beds were produced in the pluvial epoch in the Diluvium. River terraces incised in the tableland are classified by TOMITA into two groups. Like the tableland gravels, the higher terraces are capped by lateritic soil at the top. Their relative height to the present river floor is generally 100 to 300 m., but attains 400 to 500 m. on the east side of the backbone range. This difference reveals an asymmetrical geanticline caused by the upheaval of the range. The relative height of the lower terraces without lateritic soil on the top is generally 20 to 40 m. and attains 80 m. at the maximum, but no difference of height is recognizable between the two sides of the range.

While the geanticlinal upheavals were repeated, there were eruptions of dacite in the Kiirun subzone and of andesite in the Toen subzone. Because erosion is more advanced in the former than in the latter, the dacite eruption of Chinkuashih (金瓜石)-Juifang (瑞芳) must be a little older than the andesitic one of Tatunshan-Kuanyinshan (大屯山, 観音山). The early tuff of Kuanyinshan wedges into the lower part of the tableland gravels, while the foot of this volcano is covered by the gravels of the upper part. Therefore it is certain that the volcanoes appeared sometime in the Diluvium. Subsequent to the spreading of the tableland gravels the fault-angle basin of Taipei was

brought about. It was an embayment at the beginning because marine shells are contained in the lower part of the basin deposit.

In Central and South Taiwan there is a wide coastal plain and marine shells and foraminifers are found at many places in the Alluvium. The plain is widest near Taichun. There, marine subfossils are found within 10 meters below the land surface near the eastern periphery of the plain. Therefore it was a near past that the strand line ran along the foot of the hilly land. The retreat of the sea is vindicated by the historical record that a castle built at Anping (安平) in 1632 has been on a small island, notwithstanding that the locality situates 2 km. inland from the present shore line.

According to YABE the drowned valley of the Hsiatanshuichi (下淡水溪), 329 fathoms at the deepest, carves the Ryukyu limestone. Assuming this dating to be correct, the subsidence of the sea bottom since the Diluvium is not a small amount. The Pescaror trench may be the axis of the subsidence which is reciprocal to the upheaval of Taiwan.

Finally it must not be overlooked that the time length of the Diluvium inclusive of the Alluvium is no more than a million years which is, as discussed elsewhere (KOBAYASHI, 1944–45), the time-length involved in an instant in the pre-Diluvium history. If this time relation in the geologic history is considered, it is not surprising that present Taiwan is not yet far removed from the proxysm of orogeny. Taiwan is in fact still in labile state as shown by the repetition of disastrous earthquakes. Among the earthquake faults produced by the Kagi earthquake, 1906, central Taiwan earthquake, 1935, and the Tainan earthquake, 1946, all in West Taiwan, it is a remarkable habit for the northern block to shift to the east along a hinge fault in the eastnortheast trend and the downthrow of the fault is always on the south side in the east but on the north side on the westside.

Before closing this paper a few problems are discussed from the comparative tectonic point of view. The Taiwan geosyncline agrees best with the Palaeo-Shiranuhi geosyncline in Western Kyushu in the tectonic position not only at the western end of a festoon arc, but also in the inner side of the Sakawa axis. Both of them are therefore subgeosynclines, namely a minor subsiding zones which were produced secondarily in the already oronized area by its destruction. (See Fig. 8).

It is certainly interesting to see the migration of the subgeosyncline

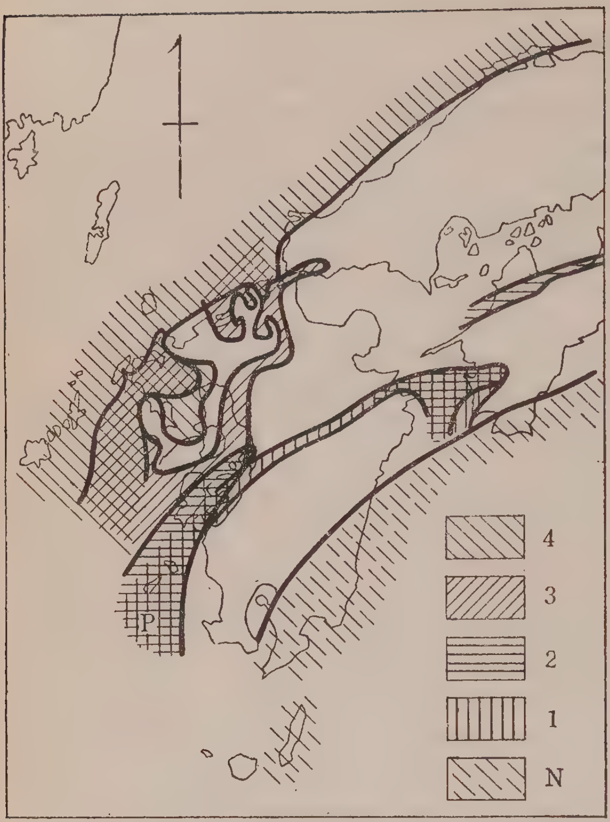


Fig. 8. Palaeographical map of Kyushu showing the migration of the sedimentary basin behind the Sakawa folded mountains.
1. Turonio-Cenomanian basin 2. Senonian basin
3. Palaeogene basin
N. Pre-Miocene Nakamura geosyncline
P. Palaeo-Shiranuhi subgeosyncline.

toward the continental side which took place in northwestern Kyushu as in Taiwan. More precisely, the oldest of the post-Sakawa depression in Kyushu was the Gyliakian or Cenomanio-Turonian one; the next the Urakawan or Senonian one; and the third the Palaeogene one. The sea ingressed toward the northeast repeatedly through the Palaeo-Shiranuhi bay, but in the inundation phase the flooded area in farther beyond was shifted from time to time toward the northwest. The shifting in this trend took place at a bound in the middle Oligocene, that is, a little prior to the paroxysm of orogeny in the Nakamura geosyncline. Through this upheaval the hinterland of this geosyncline emerged and the Oligo-Miocene sea flooded only on the northwestern periphery of Kyushu.

Sometime in the middle Neogene the Neo-Cretaceous and Palaeogene formations in the Palaeo-Shiranuhi subgeosyncline was strongly folded as the Tertiary formation in the Taiwan subgeosyncline, but the deformation of the Tertiary formations in well consolidated northern Kyushu was a kind of *Bruchfaltung*.

Because I have discussed the linking of the Japanese arc with

the Ryukyu in a paper presented to the VIII Pacific Science Congress, I do not intend to reiterate here. But I wish to call attention to the fact that the deformation of the Tertiary formations in the Ryukyu arc is not strong except the pre-Miocene Kumage in the northern part of the outer zone. In the southern part of the axial zone there is the Eocene Miyara formation comprising the *Pellatispira* limestone. It is a nearshore sediment and is overlain by andesite flows. The Miocene formation which overlies the older formations and rocks including the tilted Palaeogene is composed of andesite, andesitic tuff and agglomerate in the lower part and coal-bearing sediments in the upper part. The facies is quite different from the neritic and non-volcanic Miocene Kukinaga on the folded Kumage formation. (See Fig. 3).

The Pliocene Shimajiri formation is extensive in the middle part of the axial zone, but absent not only in the west of Miyako-jima, but also in Tanegashima and other northern islands. Therefore the distribution suggests an ingression rather than a transgression. The Sonai conglomerate is the equivalent to the Shokkozan conglomerate and occurs in limited areas of the Yaeyama group.

The Ryukyu limestone is extensive in the Ryukyu islands as suggested by its name. The eroded flat-plane on the limestone is covered by the Kunigami gravels, the correlative of the tableland gravels in Taiwan. *Palaeoloxodon* and other mammalian remains found in fillings of fissures and caves of the Ryukyu limestone indicate the latest connection of the islands with the continent which seems most probable to have been maintained at the neck of the Palaeo-Taiwan peninsula. The land bridge was, however, destroyed at length probably by the southern advance of the Ryukyu islands relative to Taiwan, yielding a flexure between them.

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Succodium, a New Codiacean Genus, and its Algal Associates in the Late Permian Kuma Formation of Southern Kyushu, Japan

(*Studies on the Paleozoic Marine Algae of Japan - 2*)*)

By

Kenji KONISHI

(With Plates I–II)

Introduction and Acknowledgments

Three Solenoporaceans (*Solenopora texana*, *Parachaetetes lamellatus*, sp. nov. and *P.*(?) sp.), two Codiaceans (*Succodium multipilularum*, gen. et sp. nov. and *S.*(?) *undulatum*, sp. nov.) and three Dasycladaceans (*Mizzia velebitana*, *Epimastopora* sp., and *Macroporella* (?) sp.) from the Upper Permian Kuma formation of Southern Kyushu are described here. *Succodium* is a new genus instituted through this paper.

The Kuma formation was selected in 1953 by KANMERA for the type formation of the Upper Permian of Japan. The writer carried on the study with the hope to learn about its algal contents, and could distinguish these eight forms including five new ones. He expected that any evidence on the correlation between the Kuma formation of Japan and the "Upper Permian" *Bellerophon* limestone of the Mediterranean province may be obtainable, but it was in vain.

The material dealt herewith consists of about 350 thin sections of limestone collected from eight limestone lenses in the Kuma formation, prepared by KANMERA for his study of foraminifers, and deposited in the Geological Department, Kyushu University.

The writer is indebted to Mr. Kametoshi KANMERA, assistant professor of Kyushu University, for the privilage of studying his valuable collection. Sincere thanks are due to Professor Teiichi KOBAYASHI of the University of Tokyo, for his encouragement and supervision of this study. The writer is grateful to Professor Seibin ARASAKI of the University of Tokyo, and Professor Yukio YAMADA of the Hokkaidô University, for their advices from the botanical side. Thanks are also extended to Messrs. Chûzaburo UEKI and Shirô SUZUKI for

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their preparation of the illustrations.

The Occurrence

The Kuma formation is an orogenic deposit, more than 900 m thick, composed of conglomerate, shale, and sandstone, as outlined from KANMERA's paper (1953) as follows; in the formation, limestones occur in three to four horizons, which are always intercalated within conglomerate or shale beds, as biostrata, lenses, or fossil banks in irregular form. In general, a lens is 3 m thick and less than 10 m long, but attains about 15 m of thickness and 25 m of length. The limestone is black to blackish gray, sandy to muddy, and so impure that rounded granules of various rocks are contained in it. Microscopically, it is mainly composed of aggregates of various fragmentary fossils, and fragments of limestone containing fossils. It is generally simillar to that of conglomerate in texture, containing rounded granules of igneous rocks, chips of black shale, and various mineral grains.

Thus, most limestones of the Kuma formation may be a kind of calcirudite.

As dendroid algae are dismembered immediately after death, it is sometimes difficult to figure out their complete forms and to make specific identification (e.g. Pl. I, Fig. 13). It is especially so when they were widely transported by current and mechanically disintegrated, abraded, and mixed with various other fossils. Except *Succodium multipilularum* in a limestone fragment[1], almost all of the fossil algae are in calcirudite in which "pieces of fossils" are intermingled with fragments of rocks and minerals. The outer parts of these fossils are generally broken or abraded[2].

The Kuma florule consists of eight forms (Table 1), each represented by ten fragments or less, except *Epimastopora* sp. Therefore fossil algae are far less in amount than other organisms (e.g. foraminifers). Assuming that the specimens were sampled at random,

1) Strictly speaking, fossils like this species in a limestone fragment are a sort of "derived fossils" brought from a pre-existed limestone by erosion. According to KANMERA (oral communication), however, fusulinids from a limestone lens at a locality consist of synchronous species and there is no older one. Therefore he is of opinion that they were derived by "contemporaneous erosion".

2) Solenoporaceans, especially nodular ones, as they are well round, are generally better preserved than dendroid algae.

two opposite explanations come to mind for the scarcity of fossil algae. One is that calcareous algae declined phylogenetically in the Late Permian or biogeographically in Kyushu or the other is that they flourished but were not preserved as fossils due to unsuitable environment for preservation. Judging from their associates, sedimentary facies, and also the number of algal species, however, the former interpretation is not tenable. Paucity of fossil algae is better explained by the second interpretation, because of unsuitable environment of the Kuma formation as suggested by the calciruditic texture[1] and

Table 1.

Locality (Number of slides)	Lower (2nd Horizon) 442 (100)	474 (13)	499 (150)	138 (7)	104 (46)	169–b (4)	Upper (3rd Horizon) 70 (15)	165 (14)
Epimastopora sp.	×		×				×	×
Macroporella(?) sp.			×					
Mizzia velebitana SCHUBERT			×					
Succodium multipilularum, gen. et sp. nov.			×					
S. (?) *undulatum*, sp. nov.		?	×					
Solenopora texana JOHNSON	×							
Parachaetetes lamellatus, sp. nov.	×		×					
P. (?) sp.	×		×					
Porostromata	×				×			

Kawamata area

Ku442; Nakagawara-dani, Kawamata-mura, Yatsushiro-gun, Kumamoto Pref.

Ku474; Northern end of a small valley running from between Kasamatsu (笠松) and Sômi (早水), Kawamata-mura, ,, ,,

Ku499; Ca. 1.5 km north of Sômi (早水), Kawamata-mura, ,, ,,

Kakisako-Kuriki area

Ku104; Northern small valley running from between Kawaiba (河合場) and Hitotsu-uji (一ッ氏), Kakisako-mura, ,, ,,

Ku70; Just south of Kawaiba, Kakisako-mura, Yatsushiro-gun, ,,

Ku138; Futae (二重), Kakisako-mura, ,, ,,

Ku165 & Ku169–b; Shiro-tani, Kuriki-mura, ,, ,,

1) A significant characteristic of the calcirudite lies in plenty of limestone fragments mainly composed of sponge spicules.

because of greater difficulty of fossilization for algae than for other organisms such as foraminifers, bryozoans, corals, etc.

The Kuma Florule (See Table 1)

Mizzia velebitana appears in Japan in the upper subzone of the Zone of *Pseudoschwagerina*. Its occurrence in the Kuma formation proves its long life range. The occurrence of *Epimastopora* sp. also shows a long range for the genus, contrary to its restriction to the Pennsylvanian and Wolfcampian. *Solenopora texana* was first described from the *Polydiexodina* Zone of the Apache Mountains which is a little lower than the niveau of the Kuma formation. Among the new forms, *Parachaetetes lamellatus* is the first Permian species of the genus and may be related to *P. triasinus* (VINASSA). But, as noted by PIA (1939), the Solenoporaceae may bear no value for wide correlation. In view of peculiar structures of *Succodium multipilularum*, its rôle as an index fossil seems promising. *S.*(?) *undulatum* tentatively referred to the new genus is also an unusual alga. But this and remaining two forms may be less important for Permian biostratigraphy than the above noted ones.

At any rate, the Kuma florule is a representative of marine algae from the Zone of *Lepidolina* or the upper subzone of the Zone of *Yabeina*, assuming KANMERA's correlation to be correct.

Finally the absence of *Gymnocodium*, the diagnostic "Late Permian" genus (PIA, 1930–35 et al.), in the Kuma florule is a remarkable fact. It is quite probable that the absence means the extinction of *Gymnocodium* before the age of the Zone of *Lepidolina*. Accordingly it is not improbable that the Zone of the Circum-Pacific is younger than the "Upper Permian" *Bellerophon* limestone of the Mediterranean province rich in *Gymnocodium*.

Systematic Description

Class Chlorophyta, Order Dasycladales
Family Dasycladaceae KÜTZING, 1843 em. HAUK, 1884
Genus *Mizzia* SCHUBERT, 1907
Mizzia velebitana SCHUBERT
Plate I, Fig. 9

1954 *Mizzia velebitana* KONISHI; *Japan. Jour. Geol. & Geogr.*, Vol. 25, p. 4 (Synonymic references given).

A transverse section of this species assures its survival in the Zone of *Lepidolina*. In addition the writer found it at other localities of the same Zone (at Dôdô, Kitawake-mura, Katsuta-gun, Okayama Prefecture (KONISHI, 1952) and Yata, Irie-mura, Kasa-gun, Kyoto Prefecture (Loc. KAMBE 10470116 in Coll., Geol. Inst., Univ. Tokyo)). As shown in the table, the Kuma form differs from those from the above mentioned localities simply in smaller diameter of verticillatae.

Measurements (in mm)	D	d	s	p	w
Ku499–3(GK–Q1005)	1.612	1.196	0.1962–.2289	0.1308–.1471	20+?

Genus *Epimastopora* PIA, 1922

Epimastopora sp.

Plate I, Figs. 2 & 3.

This is ubiquitous through the Kuma localities. Statistical data on four out of about 15 thallus-fragments belonging to the genus are given and two selected specimens illustrated. The Kuma form is closely allied to *E. kosakiensis* KONISHI from the Middle Permian Kosaki formation (KONISHI, 1954), though different in age. It is the latest survivor of *Epimastopora* so far known.

Measurements (in mm)	l	b	p	i
Ku442–32	2.964	1.040	0.0981–.1635	0.1200–.1471
Ku442–67(GK–Q1003)+)	1.716	0.520		
Ku499–29(GK–Q1004)+)	1.664	0.213	0.0654–.0817	0.1200–.1300
Ku499–147	1.194	0.180	0.0654	0.1144–.1308

+) Illustrated specimens.

Genus *Macroporella* PIA, 1912

Macroporella(?) sp.

Plate I, Figs. 5 & 12

Thallus cylindrical, more than 2.0 mm long and 1.5 mm wide with uncalcified central stalk, 0.94 mm thick. Branches simple, almost uniform in thickness (0.0327 to .0490 mm) and perpendicular to the surface. Vertical interspaces among them 0.0572 to .0684 mm. Calcified wall 0.29 to .34 mm thick.

The description is based on two fragments in longitudinal section in Fig. 5 (Ku499–131) and Fig. 12 (Ku499–110), of which the upper part or one side of the fragments is unpreserved. This form may belong to a new genus, instead of being a *Macroporella*, because branches in the latter are almost always inclined against the principal axis of thallus and their thickness increases towards the surface.

Order Siphonales

Family Codiaceae KÜTZING, 1843 em. HAUK, 1884

Genus *Succodium*, gen. nov.

Genotype;–Succodium multipilularum, sp. nov.

Generic diagnosis;–A lime-encrusting codiacean of dendroid form (*Stämchentypus*), composed of articulating segments; each segment terete with rounded ends, 2 to 3 mm long and 0.5 to 1.0 mm wide, composed of three components, namely (1) a feebly calcified central medulla of longitudinal, ramified medullary filaments, (2) a conspicuously calcified subcortical part of irregularly interwoven utricles, and (3) a thin outermost cortical layer outlining outlets of tapering utricles; gametangia-like expansions densely disposed in the same level, in forming the boundary between the subcortical and cortical parts.

Text-fig. 1. *Succodium multipilularum*, gen. et sp. nov. Longitudinal section of internode. (ca. ×40)

Occurrence;–Kuma formation, Late Permian (Zone of *Lepidolina* or upper subzone of the Zone of *Yabeina*) in Japan.

Discussions;–Succodium can safely be referred to the Codiaceae, because of development of tangled and ramified medullary filaments and utricles, both being the characteristics of the family. The most remarkable peculiarity of the new alga is the abrupt dilation of utricles to form pilules near the surface like gametangia of the Dasycladaceae. Because the pilules are disposed fairly regularly in a subsurface layer, they can hardly be mere constrictions of utricles as com-

monly seen in the Codiaceae. Although little is known of reproductive organs of the family and especially of its fossil members, the pilule reminds one of a sort of such an organ. Let us compare it to a reproductive as well as a somatic organ.

It has been generally known that reproduction of the Codiaceae is sexual and anisogamous, while the Chaetangiaceae is oogamous (EGEROD, 1952, p. 375; IYENGAR, 1951 in SMITH et al.). Among fossil dendroid codiaceans and their allies, *Gymnocodium* is a sole exception which always bears remains suggestive of a reproductive organ. But, as mentioned by PIA (1937), the remains in *Gymnocodium* are more likely conceptacles of the Chaetangiaceae than gametangia of the Codiaceae and its reference to the latter family seems very doubtful. The conceptacles of *Gymnocodium* are much larger (0.15 to .27 mm in diameter in *G. bellerophontis* from Japan) and are located in the calcified cortex more irregularly than the pilules in question.

Among living codiaceans, their reproductive organs are known of three genera[1], i.e. *Halimeda*, *Codium*, and *Avrainvillea*. In *Codium*, special ovoid or fusiform gametangia are produced within the cortex as lateral outgrowth from utricles. Utricles and gametangia are similar in their mode of attachment, but, if the difference of dimension between gametangia and utricles in *Codium* is considered[2],

	Medullary filaments	Utricles	Gametangia	
			Diameter	Length
C. arabicum KÜTZING	20–30μ	60–145(–215)μ 45–90(–150) 145–300	70–130μ	180–300μ
C. spongiosum HARVEY	30–100	85–350(–850)	60–175	215–360
C. mamillosum HARVEY	40–90	(400) 450– 1,250(1500)	230–290	600–680
C. Reediae SILVA	24–46	130–230 (85–)130–330 (–580)	80–130	260–330
C. edule SILVA	13–33	70–160 240–580 70–145 70–215	60–85	230–275

1) According to ERNST (1904), other reproductive organs of *Udotea* and *Penicillus* are really epiphytic organisms (FRITSCH, 1935, p. 413).

2) Shown in some species of the genus from Hawaii, for example, diameter of gametangia is always smaller than that of utricles. (EGEROD, 1952)

they are just the reverse in *Succodium*, namely, the pilules are larger than the utricles. *Halimeda* has a peculiar reproductive organ which is numerous spheric outgrowths[1] on its calcified thallus. This difference is quite distinct, although some structural resemblances of somatic organization can be recognized between *Halimeda* and *Succodium*. In *Avrainvillea*, "sporangia"[2] are brought forth externally at the ends of utricles (HOWE, 1907). Thus, so far as the writer is aware, *Succodium* is quite different from known codiaceans in reproductive organs.

Succodium is also discriminated by articulating segments and gametangia-like expansions from *Pseudocodium* (WEBER VAN BOSSE, 1896) which somatically resembles closely *Halimeda* rather than *Codium* but its fertile part has not yet been found. Utricles of this modern non-lime-secreting genus *Pseudocodium* are so firmly adherent together and swell up distally that they form a honey-comb structure.

Comparisons;–As discussed above, *Succodium* bears certain quite peculiar characters, but it may not be superfluous to compare it with fossil dendroid codiaceans and their allies[3]. *Dimorphosiphon* HøEG (1927) which is the oldest of the family, has thicker (200–300 μ) and less tangled medullary filaments than this genus. It is also easily distinguishable from *Anchicodium* JOHNSON, a Permo-Carboniferous genus, with the inarticulate and much slender appearance of thallus and less differentiated organization of *Anchicodium*. *Hikorocodium* ENDO (1951), the other codiacean of the same age, has generally more irregularly undulating and constricting internodes, and far larger size of utricles than *Succodium*. Distal ends of the utricles are blinded in *Hikorocodium*. In Permian, there is *Gymnocodium*, whose taxonomic position is still in question (p. 231). The abrupt expansion of the utricles as mentioned above, cannot be seen in the genus. Therefore, distinction between cortical and subcortical parts is obscure, and the

1) HOWE (1907) described diameter of gametangia of *H. tridens* to be 0.20–.33 or 0.20–.38 mm on two specimens.

2) They are 0.35 to .83 mm long and 0.2 to .35 mm wide in *A. nigricans*, which has utricles of about 0.030 to .070 mm. (HOWE, 1907)

3) According to JODOT (1935), *Microcodium* GLÜCK, 1912, may be not codiaceans (might be not organisms).

The other fossil representative of the family, *Ovulites* LAMARCK, 1816, which, according to MORELLET and MORELLET (1939), is synonymous with the modern genus *Penicillus* LAMARCK, 1813, has segments resembling those of dasycladaceans in having uncalcified medulla.

utricles are generally more tangled than *Succodium*. Possession of conceptacular remains in *Gymnocodium* is also a noteworthy distinction. There is no record of dendroid codiaceans[1] from later periods, but *Boueina* TOULA is known from the Cretaceous and *Halimeda* LAMOURAX from the Tertiary and later age, both resembling *Succodium* closely. The genus is more similar to *Halimeda* than *Boueina*, but has much slender segments and thinner medullary filaments than the second genus. *Boueina* has usually an ill-defined cortical layer. *Succodium* differs from them in the development of pilules.

Thus *Succodium* is quite unique among fossil dendroid codiaceans. Although the pilules are not readily proven to be remains of gametangia, it appears to the writer most probable that *Succodium* is a codiacean preserving such an organ[2]. It is a new genus, important as a link in the phylogeny of the Codiaceae. The generic name means a codiacean near *Codium* (sub+*Codium*).

Succodium multipilularum, sp. nov.

Plate I, Figs. 1 & 11; Text-fig. 1

Description;–Thallus dendroid, composed of articulating segments; each segment cylindrical, 2 to 3 mm long and 0.5 to 1.0 mm wide and terminated with rounded ends; medulla feebly calcified, 0.3 to 0.5 mm wide and composed of tangled medullary filaments; medullary filament, 0.025 to .033 mm thick, irregularly dichotomized and sometimes constricted; subcortical part strongly calcified, where utricles at first issue from medullary filaments almost rectangularly, then ramify and become intricated. Still later they become perpendicular to the surface in the lower one-third of a segment, but inclined at about 70 to 80 degrees to the axis of a segment in the upper two-thirds. Diameter of utricles most uniform throughout the subcortical part and similar to that of medullary filaments; the utricles abruptly expand to form pilules which correspond probably to gametangia; gametangia spheric in the proximal half, but cone-shaped in

1) In Jurassic, there is *Stenogrammites* founded by KRETSCHETOVITSCH (1936) as a dendroid red alga, which, however, lacks lime-encrustion and also such a fine organization as seen in *Succodium*.

2) As the number of specimens is small and the specimens examined are represented by only "fertile" ones, any distinction between fertile and sterile states of this alga cannot be noted here.

distal half where they terminate at minute outlets on the surface; centers of the gametangia located about 0.08 mm inside the surface.

Measurement of the Holotype;—

(in mm)	L	D	d	dmf	du	dg
Ku499–47(GK–Q1001)	2.236	0.8502	0.3107	0.0245–.0327	0.020–.033	0.0376–.0483

Occurrence;–Loc. Ku499. Preserved in limestone fragment of calcirudite (See foot-note[1] on p. 226).

Succodium(?) *undulatum*, sp. nov.

Plate I, Figs. 6, 7 & 8

Description;–Thallus dendroid, irregularly constricted at some points to form nodes; internodes cylindrical, partly undulated, about 1 mm in average breadth; medulla 0.5 mm or more broad, usually less calcified than cortex and not clearly defined; cortex segregated into two cortical parts; both defined by very small expansions of utricles at the same distance from surface; thin outer cortical part, 0.0654–.0817 mm thick; thick inner subcortical part of dichotomizing and irregularly tangled utricles which give off from medullary filaments of 0.018–.030 mm diameter.

Measurements;—

(in mm)	L	D	d	du	Th. Cortex
Ku499– 8	2.340+	1.196+		0.0213–.020	0.0659–.0817
Ku499– 16(GK–Q1002)	2.340+	1.248	0.624–.728	0.018 –.030	0.0654–.0817
Ku499– 22(GK–Q1002)	2.288+	1.404	0.468	0.0164	0.0654
Ku499– 27		0.988		0.0213	0.0490–.0654
Ku499–126		0.670	0.262		

Remarks;—As noted by HφEG (1927), PIA (1936), and others, it depends upon the state of preservation, whether or not one can detect the outermost cortical layer. The thin cortical part as seen in *Succodium multipilularum*, and *S.*(?) *undulatum*, is the exception, if species of *Halimeda*, and "*Anchicodium japonicum* ENDO (1953)[1] are excluded. Because the outermost cortical layer in the species must

1) So far as his three illustrations are examined, "*Anchicodium japonicum* ENDO" seems to the writer a composite species and the specimen of his fig. 7 on pl. XII may be separable from the species. Though ENDO gave no statement on the cortical part of the above species, two illustrations (fig. 5 on pl. XI and fig. 5 on pl. XII) appear to indicate a similar layer as *Succodium*.

be one of the characteristics of *Succodium*, this species is provisionally referred to the genus, though it differs from the genotypic *S. multipilularum* in the undulated and constricted outline of internodes, feebly defined medulla, smaller size of gametangia-like expansion in utricles near surface. Moreover, the specimens at hand do not show any positive evidence for articulation of internodes. Therefore, it is not improbable that this species represents a new genus by itself. *Holotype*;-(Syntype GK-Q1002) Ku499-16 & -22.

Class Rhodophyta, Order Florideae
Family Solenoporaceae PIA, 1926
Genus *Solenopora* DYBOWSKI, 1878

Solenopora PIA, 1930, 1939 etc.
Solenopora (part) of authors.
Genotype;-*S. spongioides* DYBOWSKI, 1878

List of the Permian Solenoporaceae is given below.
Solenopora centurionis PIA, 1940 (=*S.* sp. nov. PIA, 1937)
S. centurionis JOHNSON, 1942 & 1951; Zone of *Polydiexodina*
Solenopora texana JOHNSON, 1951 (=*S.* sp. PIA, 1940); Zone of *Polydiexodina*.
Solenopora ? sp. PIA, 1937 (=Corallinacean STEINMANN, 1919=*Solenopora* or *Lithothamnium* LEMOINE, 1917); Zone of *Yabeina*

Solenopora texana JOHNSON

Plate II, Figs. 1, 2 & 3

1951 *Solenopora texana* JOHNSON; *Jour. Paleont., Vol. 25*, No. 1, p. 23, pl. 6, figs. 4 & 5.

Solenopora texana was instituted by JOHNSON to distinguish it from *S. centurionis* PIA in the Apache collection; cells smaller in the former than in the latter. In the Kuma collection there are two fragments of *Solenopora* (s. str.) which coincide with original description of *S. texana*.

Measurements (in mm)	Size of thallus-frg.	Cell width longitdl. s.	Cell width transverse s.	Wall Thickness
Ku499- 78(GK-Q1008)*	2240	10	11	6
Ku499-130	1560×1200+	11-16		5
S. centurionis	1000-4000	33-48	28-39	10-15
S. texana	500-6200	6-13		

*) Illustrated specimen.

Genus *Parachaetetes* DENINGER, 1906

(=*Solenophyllum* MASLOV, 1935)

Solenopora (part) of authors.

Genotype;-P. tornquisti DENINGER, 1906

It is a moot discussion by several authors, whether the genus is valid or not. The taxonomy formerly proposed for Solenoporaceae appears to be rather artificial and confusing. After hesitation, however, the writer followed PIA (1939) in segregation of *Parachaetetes* from *Solenopora* (s. l.) for convenience. The genus ranges from Ordovician to Cretaceous. *P. lamellatus* described below is a Permian species which may be a link between lower Tournaisian *P. velbertiana* (PAUL) and Triassic *P. triasinus* (VINASSA).

Parachaetetes lamellatus, sp. nov.

Plate II, Figs. 4 & 5

Material;-The new species is based on eight longitudinal and two transverse sections of incomplete fragments.

Description;-Thallus composed of layers sinuously but rather concentrically arranged; occasionally, two subsequent layers discontinuously or irregularly overlapped in subparallel; each layer consists of parallel rows of cells with numerous partitions; vertical partition running continuously throughout lamellose layers. Cells 30 to 65 μ long, 15 to 30 μ broad. Cell wall straight, as thick as 10 to 20 μ, apparently lacks any special structure. In transverse section the structure of thallus appears to be composed of loosely arranged polygonal cells, 20 to 30 μ in diameter.

Comparison;-Though its external form is unknown, the wavy lamellose structure in the holotype reminds one of an encrusting form.

It appears similar to a stromatoporoid, but the present species holds the algal nature in much smaller cell size. The lamellate character and rather thick wall of cells are the most remarkable characteristics, through which *P. lamellatus**) is separable from most other species of the genus. In this respect, however, it is allied to the following three species.

Table 2.

	Longt. Sec.		Transv. Breadth	Cell Wall	
	Length	Breadth		Thickn.	Sinousity
P. asvapatii RAMA RAO et PIA	40–120μ	40–60			
P. compactus (BILLINGS)	70–130	40	50–85		
P. c. var. *norvegicus* (KIAER)					×
P. c. var. *peachi* (N. et E.)					×
P. gotlandicus (ROTHPLETZ)		25–30 (25–35)			
P. gracillis (GARWOOD et GOODYEAR)	25	17 (20–25)	17		
P. lithothamnioides (BROWN)	35–71	50–100			
P.(?) *ponticus* (DENINGER)	90–210		48–84 (60–75)	9–15	×
P. tornquisti DENINGER	(140–170)		50–70		
P. triasinus (VINASSA)	100–200		(70–150)		

*) Besides the eight species and two varieties with which its comparison is made, there are four to five species, whose descriptions are not available to the writer, and they are *Solenopora compacta onareanensis* FRITZ (1941); *P.*(?) *discoidalis* (LE MAITRE) from upper Domerian (LE MAITRE, 1937); *P. paleozoicus* (MASLOV) from "Etroeungtian" (MASLOV, 1935); *P. velbertianus* (PAUL) from lower Tournaisian (PAUL, 1935); *P. megalocytus* PIA from middle Jurassic (PIA, 1941).

PIA is of opinion that *Petrophyton* YABE is synonymous with *Parachaetetes*, but the writer segregates the following three to four species as the former; *P. kiaeri* HøEG, *P. miyakoense* YABE, *P. tenue* YABE et TOYAMA, *P.*(?) *rothpletzi* YABE.

P. lithothamnioides (BROWN) resembles this Permian form but differs from it in cell outline. The Ordovician form has cells of quadrate outline, elongated transversely. The second ally is *P. toronquisti* which is approximate in thickness of the cell wall but cells are larger. The new form is also akin to *P. triasinus* but distinguishable by far larger cells.

Measurements (in μ)	Size of thallus-frgm.	Cell length longitdl. s.	Cell width longitdl. s.	Cell width transvrs. s.	Wall thickns.
Ku442–12	3276	65.4	29.4		
Ku442–70(GK–Q1009)*	8800	36.1	15–24		16
Ku442–74	3016	36.1	15–24	13–16	10

*) Holotype (Illustrated).

Occurrence;–Ku442, 499, & 70

Parachaetetes(?) sp.

Plate I, Fig. 10

There are five fragments tentatively referred to a species of the genus, though their transverse partitions are sometimes not clear. The form is distinguishable from the preceding by its thinner and ill-defined cell wall, especially by obscurity of concentric partitions and also by the absence of lamellose structure. The last character suggests a nodular form. All the specimens are so poorly preserved that no precise description can be given. Besides the Kuma formation (Locs. Ku442 & 499) some fragments referable to this form are recently found from the Dôdô Conglomerate, an equivalent of the formation (Konishi, 1952).

Measurements (in μ)	Size thallus-frgm.	Cell width longitdl. s.	Cell width transvrs. s.	Wall thickns.
Ku442–57	3224 × 4680	41–49	25–33	16
Ku499–119	2288 × 2600		16–27	10
Ku499–139(GK–Q1008)*	1976 × 2964	57–68	12	10–15

* Illustrated specimen.

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* These literatures were not directly available through the present study.

Postscript; While the present paper is in press, KANMERA's paper on "Fusulinids from the Upper Kuma Formation, Southern Kyushu, Japan with Special Reference to the Fusulinid Zone in the Upper Permian of Japan (*Mem. Fac. Sci. Kyushu Univ., Ser. D, Geol., Vol. IV,* No. 1)" came out. The reader is requested to refer to it for detailed informations of the formation and its fusuline fauna.

Plate I.

Explanation of Plate I.

Page

Figs. 1 & 11 *Succodivm multipilularum*, gen. et sp. nov. 233
Holotype ; longitudinal section (Ku499–47 ; GK–Q1001) Fig. 1 ×20 Fig. 11 ×40

Figs. 2 & 3 *Epimastopora* sp. 229
Incomplete obliquely longitudinal sections ×20 Fig. 2 Ku442–67 (KG–Q1003) Fig. 3 Ku499–29 (GK–Q1009) Fig. 4 A frgment of indeterminable dasycladacean (?) ×20 (Ku442–98)

Figs. 5 & 12 *Macroporella* (?) sp. 229
Incomplete longitudinal sections ×20 Fig, 5 Ku499–131 Fig. 12 Ku499–47 (Syntype, GK–Q1006)

Figs. 6, 7 & 8 *Succodium* (?) *undulatum*, sp. nov 234
×20 Holotype (Syntype) (GK–Q1002; Ku499–16 & 22) Fig. 6 Ku499–22–22 Longitudinal section Fig. 7 Ku499–16 Longitudinal section Fig. 8 Ku499–126 Cross section

Fig. 9 *Mizzia velebitana* Schubert 228
Transverse section ×20 (Ku499–3) (GK–Q1005)

Fig. 10 *Parachaetetes* (?) sp. 238
Obliquely transverse section (Ku499–139) ×40 (GK–Q1008)

Fig. 13 Fragments of indeterminable dasycladaceans........................ 226
×20 (Ku499–35)


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Plate II.

Explanation of Plate II.

Page

Fig. 1, 2 & 3 *Solenopora texana* Johnson 235
Fig. 1 ×100 Fig. 2 ×40 Both (Ku499–78; GK–Q1008). Transverse Section

Figs. 4 & 5 *Parachaetetes lamellatus*, sp. nov. 236
Fig. 4 ×100 Fig. 5 ×40 Both (Ku442; GK-Q1009)

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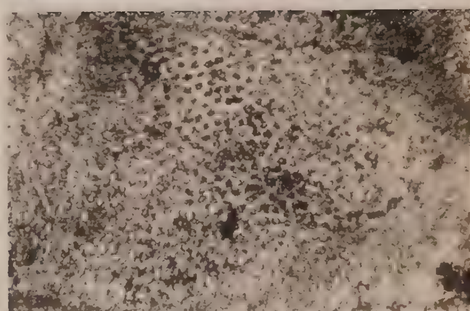

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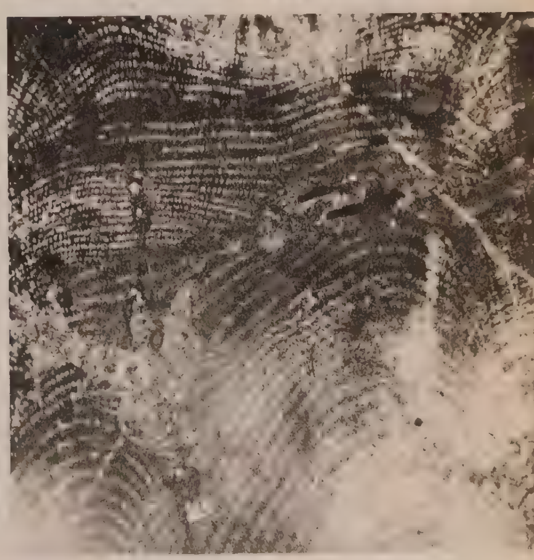

5

Geology and Petrology of Ōmuro-yama Volcano Group, North Izu

By

Hisashi KUNO

Abstract

Ōmuro-yama Volcano Group comprises small cones, domes, and pyroclastic layers erupted from more than ten separate vents. The age of eruption ranges from younger Pleistocene to Recent. The rocks of the volcanoes vary in composition from olivine basalt to andesite. Most of them contain xenocrysts captured from a granitic rock which constitutes the wall of the magma reservoir underlying the volcanoes. The depth of the magma reservoir is estimated as more than several kilometers from the surface. The Itō earthquake swarm of 1930 was caused by intrusion of a part of the magma still surviving in the reservoir. Most of the rocks of the Ōmuro-yama Group are dissimilar to those of Ō-sima Volcano lying 30 km southeast of the former, being higher in $MgO : FeO+Fe_2O_3$ ratio and higher in alkalies. This difference is interpreted as due to contamination of the magma of Ōmuro-yama by the granitic rock.

Introduction

The main scope of this paper is to describe the geology of Ōmuro-yama Volcano Group lying on the eastern coast of north Izu Peninsula between 139°01′ and 139°09′ E and 35°0′ and 34°48′ N. This group has a unique structure among the volcanoes of the Huzi Volcanic Zone. Thus, instead of forming a large cone as in the other volcanoes, it consists of many small volcanoes erupted from separate vents. The ages of eruption of the individual vents were also determined.

In this paper is also discussed the relationship between the seismic activity of 1930 and the subsurface magmatic activity related to this volcano group.

Another topic of this paper is the petrographic comparison with the rocks of the volcanoes of the southern part of the Huzi Zone, especially with those of Ō-sima which is the nearest active volcano.

In 1920, Dr. Seitaro TSUBOI published an excellent paper on the geology and petrology of the Ō-sima Volcano. He found that the basalt lavas of Ō-sima, especially the historic ones, are characterized by low $MgO : FeO+Fe_2O_3$ ratio and low alkalies, and also by normative

feldspar rich in An and high normative quartz. Since his finding, similar rocks have been reported from adjacent volcanoes (TSUYA, 1937; KUNO, 1950 a) and are now generally accepted as constituting the bulk of the volcanoes of the southern subzone of the Huzi Volcanic Zone (KUNO, 1952 a).

The rocks of the Ōmuro-yama Group which lies a little to the west of the trend of the southern subzone have different characters, being high in $MgO : FeO + Fe_2O_3$ and high in alkalies. Some of them have a considerable amount of normative olivine while normative feldspar is not so high in An. Most of them contain resorbed xenocrysts captured from a granitic rock.

The rocks of the southern subzone are explained as representing various stages of fractionation of the tholeiitic magma (the pigeonitic rock series of KUNO) (1950 a), while most of the rocks of Ōmuro-yama as representing products of contamination by the granitic rock (the hypersthenic rock series of KUNO) (1950 a). Only a brief discussion on this subject will be given in this paper.

The geology of the Ōmuro-yama Volcano Group was first described by J. SUZUKI (1921), and later by H. TSUYA (1930) who gave also petrographic description of the rocks and pointed out the presence of quartz and andesine xenocrysts (1930, 1937). TSUYA's conclusion as regards to the structure of the volcanoes was largely confirmed by the writer's study. The geologic map shown in the present paper is partly based on TSUYA's map.

Acknowledgments

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Geology

The Ōmuro-yama Volcano Group comprises flat shield volcanoes, stratovolcanoes, a lava dome, and layers of scoriaceous ejecta. They were erupted from more than ten separate vents (Figs. 1 and 2).

Fig. 1. Distribution of vents of Ōmuro-yama Volcano Group and epicenters of earthquakes of 1930.

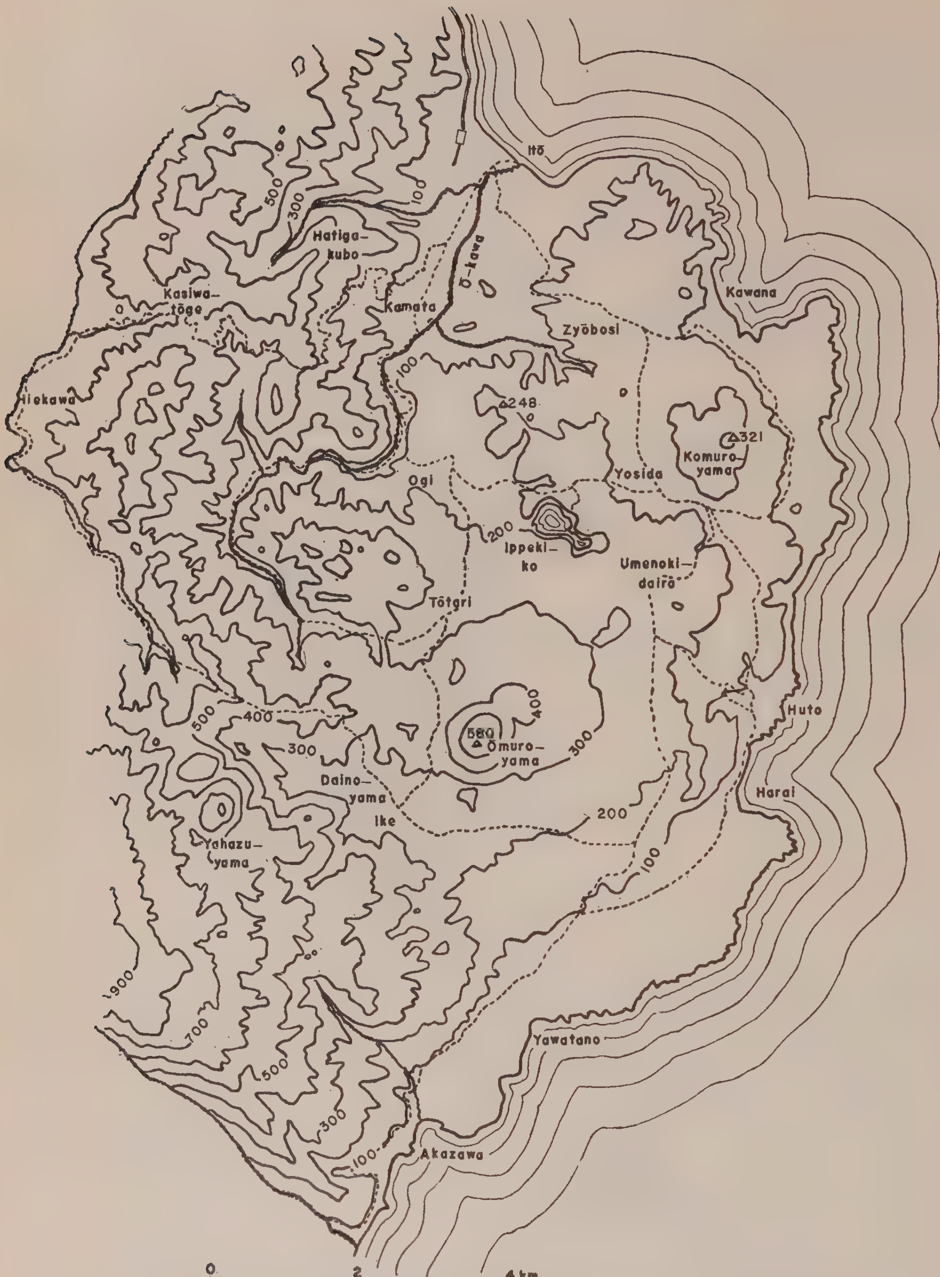


Fig. 2. Topographic map of the area of the main volcanic activity.

The area where most of the vents are clustered lies just south of the city of Itō, and is largely made up of gently sloped lava fields (Fig. 2) covered by ash and lapilli. To the west of this area rises a deeply dissected mountain consisting of older volcanic rocks, and to the southwest is a gentle slope of the Amagi Volcano which is also older than the Ōmuro-yama Group. The geologic map of these areas is shown in Fig. 3.

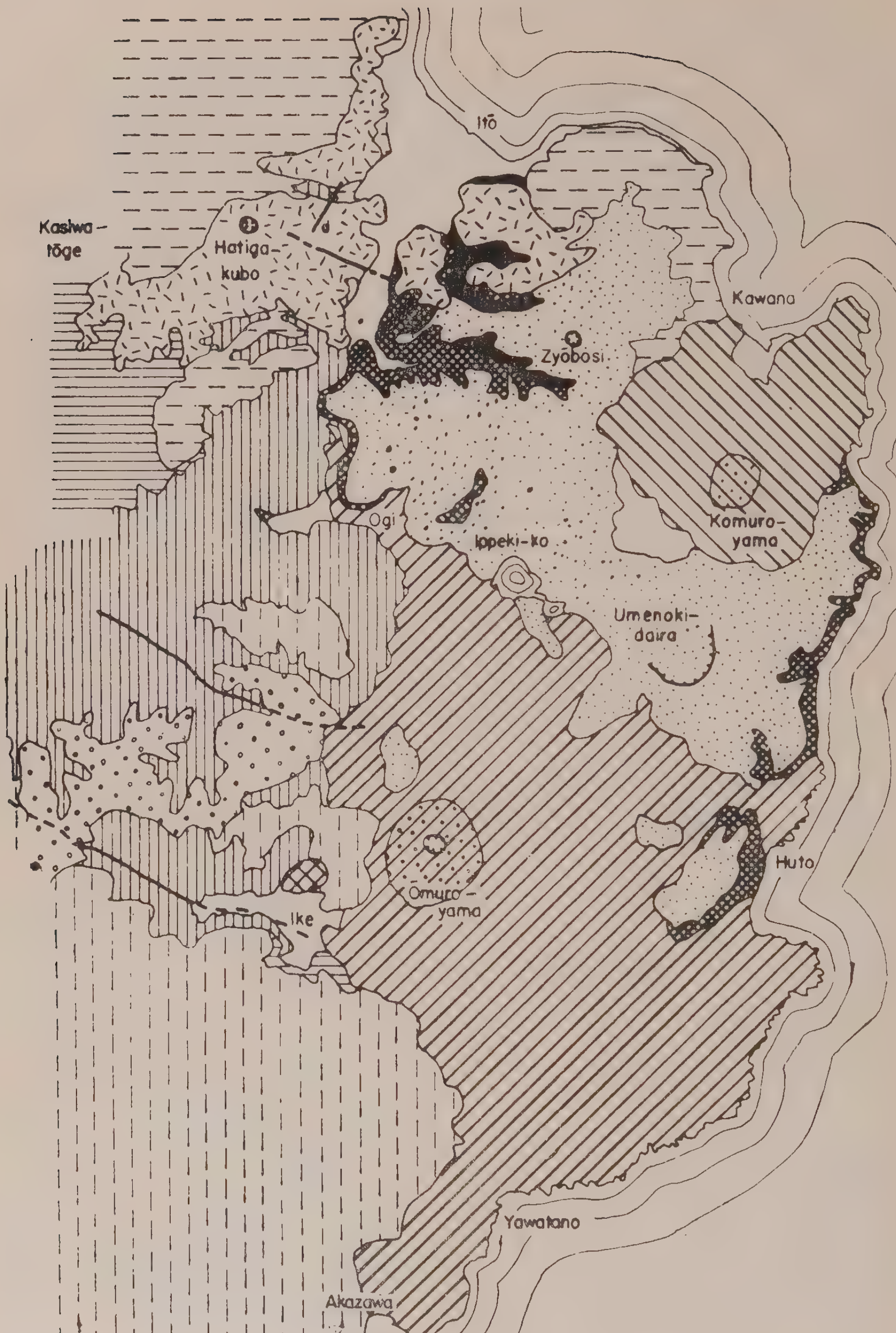
Two isolated vents are also shown in Fig. 1, Sukumo-yama on the northwest and Ōta on the south.

Basement Rocks. The oldest formation in the area included in Fig. 3 is the Yugasima Group of older Miocene age (KUNO, 1951) exposed in the middle western part of the map. It consists of altered pyroxene andesites and lapilli-tuff. The strata dip to east and north-northeast at angles about 45°. Two faults cutting through the formation are recognized. Both of them strike west-northwest. Probably this formation extends eastward and underlies directly the lavas of the Ōmuro-yama Group.

In the northwestern part of the map, the Yugasima Group is covered by volcanic rocks of younger Miocene to Piocene age, consisting of essential tuff breccia (definition by KUNO, 1951, p. 364) of andesite and lapilli-tuff of dacite, and also by andesite lavas of the Usami Volcano (TSUYA, 1937; KUNO, 1952 b) of older Pleistocene age. The lavas are also exposed to the east of the city of Itō where they dip generally southeastward and are covered by the lava and scoria of the Ōmuro-yama Group.

The center of the Amagi Volcano lies beyond the southwestern corner of the map Fig. 3. It is a large stratovolcano made up of andesites and a little basalts and dacites. The map includes only a part of its northeastern slope. Parasitic lava domes and side craters are scattered on the northeastern foot of the main cone. Basalt ash and lapilli from one of these side craters are shown in the map.

Huzi Ash. Although not shown in Fig. 3, part of the present area is covered by a layer of basaltic ash a few meters thick. The materials of this layer came largely from Huzi Volcano lying 50 km northwest of Ōmuro-yama and partly from some other volcanoes. The layer can be traced from the southeastern foot of Huzi down to the present area. The characteristic features are the dark brown color, absence of stratification, and peculiar mineral composition. The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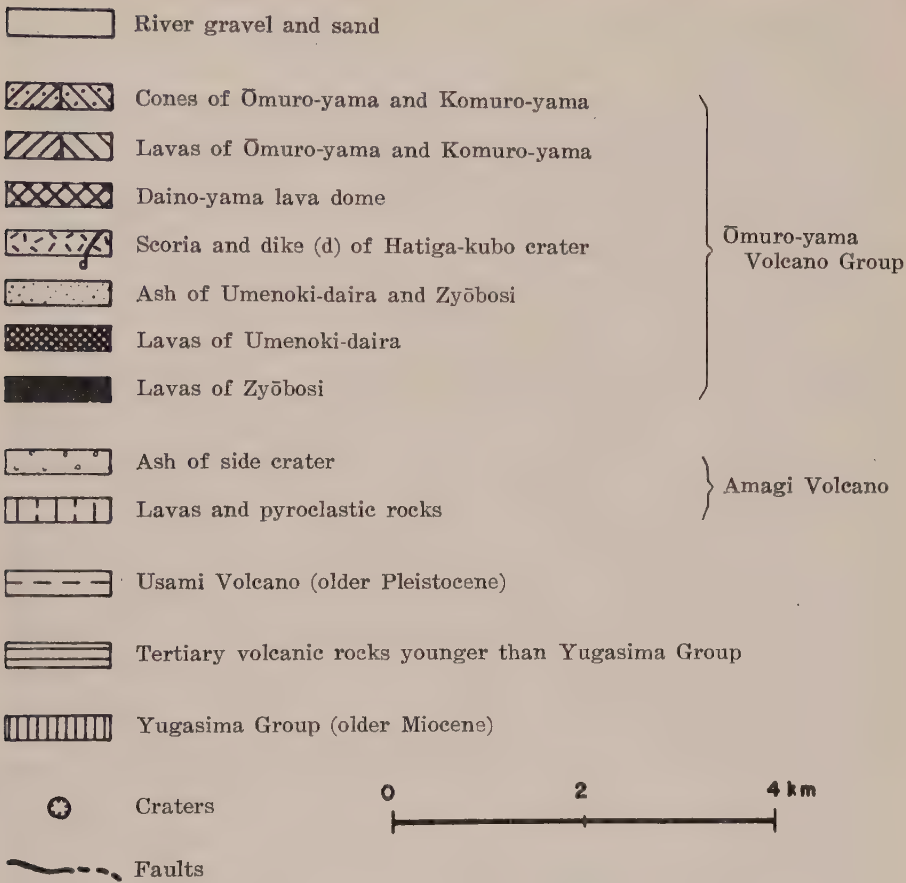

Fig. 3. Geologic map of the area of the main volcanic activity.

minerals are olivine, hypersthene, augite, magnetite, green hornblende, and plagioclase, in the order of abundance, although their proportion varies in different localities and in different horizons of the layer. Thus in the area northwest of Itō, hornblende is abundant in the lower horizon but almost absent in the upper, while olivine tends to decrease in amount toward the lower horizon. Since hornblende increases in amount toward the southeastern area, the mineral was derived from a certain volcano near the eastern coast of Izu Peninsula.

Two or three thin layers of yellow pumice of pyroxene andesites are intercalated between the ash in its middle and lower horizons. The uppermost layer consists of pumice from one of the central cones

of Hakone Volcano adjacent to the southeast of Huzi (KUNO, 1950 b, p. 273).

To the east of the Huzi Volcano, the same ash covers an extensive area, forming the Kwantō Volcanic Ash (formerly called Kwantō Loam) of younger Pleistocene age or about 10,000 years old.

The Huzi Ash overlies some of the ejecta of the Ōmuro-yama Volcano Group and underlies others.

The individual volcanoes of the Ōmuro-yama Group are described below in the general order of eruption.

Zyōbosi Volcano. A flat-lying lava of compact olivine basalt, more than 20 m thick, was extruded from a vent situated 2.5 km southeast of Itō. The location of this vent is indicated by a shallow crater named Zyōbosi, a few hundred meters across. Coarse scoriaceous ejecta and bombs are accumulated around this vent. The lava flowed down northwestward along the southwestern foot of the hill made up of the lavas of the Usami Volcano. The Zyōbosi lava is overlain by the ash from the same crater which is in turn covered by the ash and scoria of Hatiga-kubo crater to be mentioned later. It is not certain whether the lava is younger or older than the lavas of Umenoki-daira Volcano next to be described. They are probably of the same general age as judged from the degree of dissection.

Umenoki-daira Volcano. To the south of the Zyōbosi Volcano spreads a flat lava plateau (Pl. III, Fig. 2) covered by brown stratified ash and lapilli.

The lavas and pyroclastic materials were erupted from a shallow crater named Umenoki-daira, several hundred meters across, lying 6 km south-southeast of Itō. The crater wall is now represented by a semi-circular ridge convexing southeastward, forming the highest point (260 m) of the plateau. The ridge is made up of coarse pyroclastic ejecta including bread-crust bombs 2 m across. In the middle of the crater is a hill (260 m) consisting probably of pyroclastic ejecta.

The ash and lapilli of this volcano grade upward to the Huzi Ash. This relation is best seen on the road-cut between Ippeki-kọ and Ogi, west-northwest of Umenoki-daira. No significant time interval appears to exist between the deposition of the ashes of these two volcanoes.

The lavas of the Umenoki-daira Volcano consist of lower and upper members. The two are exposed on the sea-cliff just south of Huto, south-southeast of Umenoki-daira. Here the lower member is

represented by olivine-pyroxene andesite with numerous phenocrysts of plagioclase. Above this lies a flow of the upper member more than 50 m thick. It is sparcely porphyritic augite-olivine basalt.

The two members are also exposed along the northwestern to the northern margin of the plateau. Here the lower member consists of a few successive flows of porphyritic olivine-pyroxene basalt interbedded with scoria and intruded by a dike of the same rock type. The best exposure is at the southern end of the city of Itō. The upper member is represented by a flow of nearly aphyric olivine-augite basalt more than 50 m thick, exposed to the northwest of Ogi.

TSUYA (1930) mentioned the existence of a small explosion crater northeast of Ippeki-ko (marked ? in Fig. 1) from which fragments of dacite were ejected, although its topographic feature has been obliterated by recent construction of farms and roads.

There are also two hills to the southwest of the Umenoki-daira crater, one 334.5 m and another 420 m high, projecting above the lava field of Ōmuro-yama Volcano to be described later. They may represent pyroclastic cones or demolished crater walls (marked ? in Fig. 1) formed in the same general age as the Umenoki-daira crater.

Scoria of Hatiga-kubo Crater. Well-stratified, unconsolidated scoria, lapilli, ash, and bombs of olivine basalt cover the slope of the mountain southwest of Itō, and extend to the south of the same city where they rest on the ejecta of Zyōbosi and Umenoki-daira. They were ejected from a crater named Hatiga-kubo, about 100 m across and 20 m deep, lying on the slope 1.5 km west of the southern end of the city. The grain-size of the ejecta increases as we approach the crater. The maximum thickness of the ejecta layer is about 100m.

On the road-cut along the highway 3.5 km southwest of Itō, the scoria is overlain by the upper part of the Huzi Ash 1 m thick and is underlain by another ash containing hornblende, possibly representing the lower part of the Huzi Ash mixed with materials of other volcanoes. Remains of Neolithic men were found overlying the scoria at the southeastern margin of the city.

Two nearly vertical faults, both striking N 60°W and cutting through the scoria, are exposed at the eastern foot of the slope near the southern end of the city.

At the source of the water-supply for the city of Itō, lying a little east of the Hatiga-kubo crater, the scoria is intruded by a verti-

cal dike of olivine basalt more than 1 m thick, striking N 20°E. The chilled surface of the dike shows mumerous minute ribs running in a definit direction, resembling the surface feature of some basalt flows. The ribs were formed by flow movement of the magma. Thus the surface of the dike adjacent to the wall-rock was first consolidated to form a thin skin enveloping the liquid interior of the dike. As the liquid moved upward, the skin which was still a little plastic was crumpled to form the ribs parpendicular to the direction of the flow. From the general trend of the ribs it may be inferred that the magma came up from some depth to the south and moved upward at an angle of about 45° with the horizontal plane.

In a small valley just west of Itō, or a little northeast of the dike locality, a lava of olivine andesite is exposed in a small area. Possibly the lava is intercalated between the scoria.

Daino-yama Lava Dome. A dome-shaped mountain, only 100 m high from the base, lying just west of Ōmuro-yama, is a simple mass of compact, light-grey augite andesite with conspicuous xenocrysts of quartz and andesine usually 5 mm across. Because of lack of exposure, the relation with the ejecta of the other volcanoes is not known.

Sukumo-yama Volcano. This volcano lies 7 km northwest of Itō and is far removed from the area of the main volcanic activity (Fig. 1).

Sukumo-yama is a dome-shaped mountain without summit crater (Pl. III, Fig. 1). It is 580.5 m high above sea level and about 200 m high from the base. The main part of the volcano appears to consist of lavas of olivine basalt interbedded with pyroclastic layers. The exposure of one of the lavas is seen only on the northwestern flank of the mountain where it inclines northwestward at an angle of 25°. The other part of the mountain is covered by scoria, lapilli, and bombs which spread westward to a distance of 3 km from the summit.

The ejecta lie on the dissected slope of the Usami and Taga Volcanoes. In his previous paper (1952 b), the writer pointed out the occurrence of the Huzi Ash and Hakone pumice beneath the ejecta of Sukumo-yama. However, the heavy mineral analysis revealed that the pumice is not that of Hakone while the material of the ash partly agrees with that characteristic of the lower part of the Huzi Ash but has been partly derived from other sources. Complete absence of the Huzi Ash on the ejecta of Sukumo-yama also suggests that

the latter is younger than the former.

Ōmuro-yama Volcano. Nearly half of the volcanic field south of Itō is occupied by a shield volcano surmounted by a steep cone named Ōmuro-yama lying 7.5 km south of the city. At the center of the shield lies a craggy hill 430 m high (Fig. 2), representing a vent from which most of the lavas of the shield were extruded.

A little south-southwest of the craggy hill rises the cone of Ōmuro-yama which is 581.0 m high above sea level and about 300 m high from its base. It has a crater 30 m deep on the summit and a shallow crater on the southeastern flank. The cone is made up of alternate layers of lavas and scoriaceous ejecta.

A small hill on the southern foot of the cone (Fig. 2) appears to represent another vent from which some of the lavas of the shield were extruded.

The lavas of the shield volcano show a typical aa surface. The rocks are quite similar to those of the cone. They are dark grey olivine andesite with sporadic xenocrysts.

It is not certain whether the bulk of the lavas of the shield were extruded before or after the formation of the cone. Most probably the extrusion of the lavas and the growth of the cone took place almost simultaneously.

One of the lavas of the shield which flowed down southward dammed up a river, resulting in a small lake which was later drained and is now represented by the basin of Ike. In a similar way, another lake named Ippeki-ko was formed by the northern flow.

A distinct erosional unconformity is noted between the Huzi Ash and the Ōmuro-yama lavas. The latter filled up valleys which were cut down through the plateau made up of the lavas and ejecta of the Umenoki-daira Volcano and of the Huzi Ash. This relation is best seen along the road between Ippeki-ko and Ogi. The surface of the Ōmuro-yama lavas is free from covering of younger ejecta and also of weathering product. From these facts it is most likely that the lavas were extruded within the last several thousand years.

The physiographic feature and structure of the Ōmuro-yama Volcano resemble very closely those of the Parícutin Volcano, Mexico (KRAUSKOPF, 1948). Even the petrographic character of the Ōmuro-yama lavas is similar to that of the earlier lavas of Parícutin.

Komuro-yama Volcano. The physiography, structure, and nature

of the rocks of this volcano are quite similar to those of the Ōmuro-yama Volcano, although the former is a little smaller.

The lavas of the shield volcano were extruded from a vent now represented by a hill 260 m high. Komuro-yama, lying adjacent to the south of the hill, is a cone 321.2 m high above sea level and 100 m high from the base (Pl. III, Fig. 2). Owing to the lack of exposure it is unknown whether the cone is made up of pyroclastic materials and lavas or exclusively of the former.

As inferred from the absence of ash from other volcanoes and weathering product on the surface of the lavas, the volcano is probably of the same general age as the Ōmuro-yama Volcano.

Scoria of Ōta Crater. Well-stratified scoria, lapilli, and ash of olivine basalt, a few meters in thickness, cover the lower part of the slope of the Amagi Volcano near Atakawa 16 km south of Itō (Fig. 1). They were ejected from a crater now represented by a shallow depression named Ōta, a few hundred meters in diameter, lying 1 km west-southwest of Atakawa. It might be one of the side craters of Amagi and unrelated to the Ōmuro-yama Group.

Age of Volcanic Activity

The lava of the Zyōbosi Volcano, one of the oldest volcanoes of the Ōmuro-yama Group, rests on the deeply dissected slope of the Usami Volcano of older Pleistocene age (KUNO, 1952 b). From this fact and also from the relation between the Huzi Ash and the ejecta of the Umenoki-daira Volcano, another oldest volcano of the Group, it is safe to conclude that the eruption of the Ōmuro-yama Group started in younger Pleistocene.

The ages of eruption of most of the other volcanoes are also determined from the field relations with the Huzi Ash.

Table 1. The ages of the volcanoes of the Ōmuro-yama Group and of the Huzi Ash.

Recent — 5,000 years	Ōmuro-yama	Komuro-yama	Sukumo-yama		? Ōta ?
— 10,000 years	?	Huzi Ash		Hatiga-kubo	
Younger Pleistocene	Daino-yama				
		Umenoki-daira		Zyōbosi	

The ages of the individual volcanoes of the Group are shown in Table 1 in a summarized form.

Location of Magma Reservoir

As will be mentioned in later part of this paper, most of the rocks of the Ōmuro-yama Group were derived from magma which assimilated quartz diorite or granodiorite such as found as xenoliths in the lava of Ōmuro-yama. The xenoliths would have been captured from the wall of the magma reservoir which supplied the materials of the Group and not from the wall of the conduits, because, in order to effect assimilation, the magma would have been in contact with the salic plutonic rock for a considerable period of time, otherwise only mechanical mixing of the magma and the country rock would have taken place. It follows then that the wall of the reservoir was made up at least partly of the plutonic rock.

No exposures of such plutonic rocks have been found in Izu Peninsula, although they occur occasionally as ejecta in pyroclastic

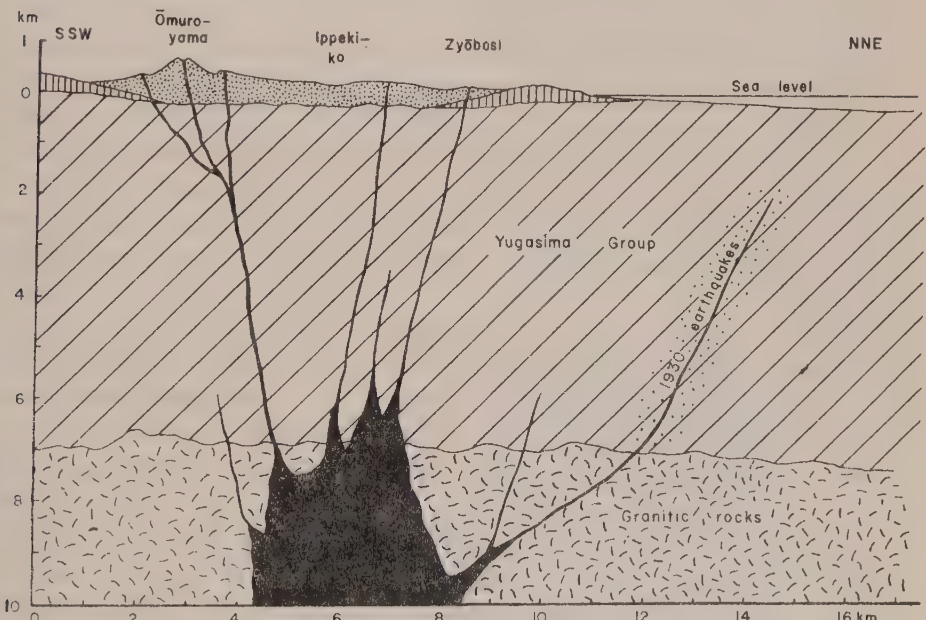

Fig. 4. Schematic profile section through the middle of the Ōmuro-yama Volcano Group in relation to the foci of the Itō earthquakes of 1930.

rocks of Tertiary age. Most probably therefore the plutonic rock underlies the Yugasima Group which forms the foundation of the Ōmuro-yama Volcano Group. As the thickness of the Yugasima Group is estimated as at least several kilometers, the reservoir was situated at a depth more than several kilometers, as illustrated in Fig. 4.

It is reasonable to consider that the location of the magma reservoir was close to the center of the main volcanic activity, namely the center of the volcanic field south of Itō. From the top of the reservoir narrow conduits extended up to the surface vents passing through the Yugasima Group (Fig. 4). Most of these conduits were nearly vertical, and the magma moved through these conduits rather quickly without causing assimilation of the rocks of the Yugasima Group.

The isolated vents such as those of Sukumo-yama and Ōta may have been connected either with the same reservoir by gently inclined conduits or with some satellitic reservoirs directly underlying these vents.

Origin of the Earthquake Swarm of 1930

The eruption of the Ōmuro-yama Volcano Group started in younger Pleistocene and continued until some period within the last several thousand years. A part of the magma may still survive in the reservoir.

During the period from February 13th to the end of April, 1930, more than 6000 earthquakes shook the area around Itō, although none of them were destructive. The locations of the earthquake foci were determined by NASU, KISHINOUYE, and KODAIRA (1931). Most of the epicenters were clustered within a limited area shown by a dotted line in Fig. 1.

This area lies on the north-northeastern extension of the belt in which most of the vents of the Ōmuro-yama Group are included. TSUYA (1930) therefore considered that the earthquakes were caused by a sort of crypto-volcanic activity of the Ōmuro-yama Group

NASU (1935) demonstrated that the earthquake foci were situated between the depths of about 7 km and 2 km, being included in a cylindrical space dipping steeply to the southwest. He pointed out a distinct tendency for the earlier earthquakes to occur at deeper horizons and the later ones at shallower horizons:—a phenomenon resembling some process of volcanic eruption. He concluded that the earthquakes were caused by magmatic intrusion.

Further evidence for the magmatic activity is furnished by the result of precise leveling along the route from a point north of Aziro, passing through Usami, Itō, and Yawatano, to Atakawa. C. TSUBOI (1931, 1933) summarized the result and showed that, during the period from the beginning of the seismic activity, or a little earlier, to January, 1933, bench marks between Itō and Yawatano had been elevated, the point of maximum elevation being situated just northwest of Komuro-yama, as shown in Fig. 5. This indicates that the area near the center of the main volcanic activity had been up-domed. As a result of continued upheaval until January, 1933, the point halfway between Itō and Komuro-yama attained a height of 340 mm above the level of the same point in 1923–24.

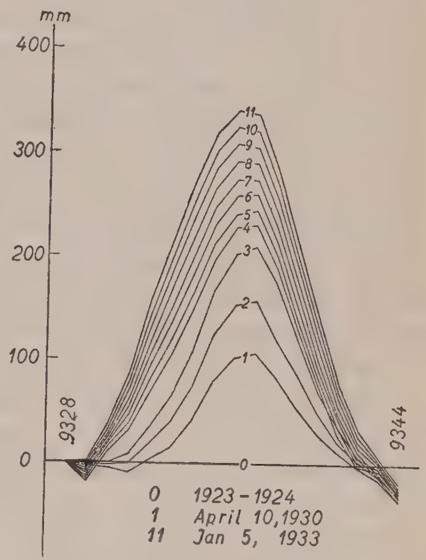

Fig. 5. Amounts of upheaval of bench marks along the eastern coast of north Izu Peninsula. Each curve represents the heights of the bench marks after every 100 days since April 10, 1930, relative to the heights of the same bench marks in 1923–24. The bench mark No. 9328 is situated 3 km northwest of Aziro, while No. 9344 4 km southwest of Yawatano. There are 15 bench marks between the two at interval of 2 km. After C. Tsuboi (1933).

According to TSUBOI, the velocity of upheaval decreased toward January, 1933. The point of maximum velocity of upheaval was first situated at a point northwest of Komuro-yama, but during and after the seismic activity it migrated gradually northward as far as to a point a little south of Usami, and then in November, 1930, when another earthquake swarm occurred in the area northwest of Usami and the destructive earthquake of north Izu occurred in the area further west, it came back again to a point near the southeastern margin of the city of Itō. However the absolute amount of upheaval of the area between Usami and Itō was not so great as compared with that of the area to the southeast, as is also seen in Fig. 5.

It is quite likely that the dome-shaped upheaval of such a small area was caused by an increase of pressure of a mass of magma.

The writer considers that the Itō earthquake swarm was caused by intrusion of a part of the magma still surviving in the reservoir. The deepest earthquake occurred approximately at the postulated depth of the reservoir (Fig. 4).

The observed distribution of the earthquake foci and the result of the leveling may be interpreted as follows:

A little earlier than the beginning of the seismic activity, the pressure of the magma in the reservoir began to increase and pushed up the roof, resulting in the up-doming of the area just above the top of the reservoir, namely near the center of the main volcanic activity. This did not cause earthquakes possibly because such a gentle up-doming was not accompanied by fracturing of the wall rocks. The magma continued to push the roof until 1933. However a part of the magma found its way to move northward, and then rose up through a narrow, steeply inclined channel, fracturing the wall rocks and causing the earthquakes. The earthquake foci were therefore arranged along or near this channel. An upward movement of magma from south to north is illustrated by the dike cutting through the scoria of Hatiga-kubo. The slight upheaval of the area between Usami and Itō may be due to the northward movement of the magma.

The magma which caused the earthquakes would have had a high viscosity, either as a result of assimilation of the salic plutonic rocks or as a result of advanced crystallization. It is suggested that the magma comparable to that of the Daino-yama Lava Dome was responsible for the earthquakes.

Petrology

The rocks of the Ōmuro-yama Volcano Group may be classified into three groups: undersaturated olivine basalt, basalts and andesite of the pigeonitic rock series, and basalts and andesites of the hypersthenic rock series.

Undersaturated olivine basalt (type III b→c according to the classification based on mineral assemblage, KUNO, 1950 a). Only the lava of Sukumo-yama belongs to this group. Skeletal crystals of olivine and tabular crystals of bytownite form phenocrysts less than 5% of the rock. Groundmass consists of tabular plagioclase zoned

from bytownite to andesine, pigeonite, augite, olivine, magnetite, and a little interstitial anorthoclase. The presence of the last-named mineral and a considerable amount of olivine in the groundmass and the absence of silica minerals are the features which distinguish this rock from the other rocks of the Izu and Hakone province. Indeed this rock is the only example in this province that yields normative olivine more than 5%, as shown in Table 2, column No. 1.

In Table 3, column No. 12, is shown the composition of the parental basaltic magma of the pigeonitic rock series of this province. This composition is obtained by averaging analyses of two olivine basalts having the most magnesian compositions among the aphyric rocks of the series (KUNO, 1953). As compared with the parental basalt, the Sukumo-yama basalt is higher in alkalies and poorer in lime; in other components they are almost identical. Normative feldspar of the former is An 78 while that of the latter is An 58.

Although the Sukumo-yama lava contains very sporadic xenocrysts of quartz, it is doubtful whether the deviation in composition from the parental basalt can be accounted for by contamination. Simple process of contamination of the saturated parental magma by a salic plutonic rock would not produce the undersaturated basalt of Sukumo-yama. Nor the difference can be explained as a result of fractional crystallization. A suggested explanation is that the difference is due to gaseous transfer or to thermo-diffusion (BARTH, 1952).

Basalts and andesite of the pigeonitic rock series. The rocks of the lower member of the Umenoki-daira Volcano and of the dike cutting through the scoria of Hatiga-kubo belong to this series.

The Umenoki-daira rocks are augite-olivine basalts (type IVc), hypersthene-augite-olivine basalt (type Vc), and olivine-augite-hypersthene andesite (type Vd→c), all containing abundant phenocrysts of anorthite-bytownite. Phenocrysts of olivine and augite are always present, while those of hypersthene are found in a few rocks. The groundmass is usually holocrystalline, consisting of plagioclase, augite, pigeonite, magnetite, and interstitial cristobalite. Xenocrysts of quartz, andesine, and hornblende are found in the andesite which contains hypersthene in the groundmass rimmed with clinopyroxene (type Vd→c).

Analysis of one of the rocks of this member is quoted in Table 2, column No. 4. Its composition agrees with those of the porphyritic rocks of the pigeonitic rock series of the Izu and Hakone province,

Table 2. Chemical compositions and norms of the lavas of the Ōmuro-yama Volcano Group.

Volcano	Sukumo-yama	Zyōbosi		Umenoki-daira	
No.	1	2	3	4	5
Type of assemblage	IIIb→c	IIIb→d	IIIa→d	Vc	IVd
SiO_2	48.10	49.08	51.82	50.17	53.13
Al_2O_3	16.68	17.65	15.90	19.30	16.57
Fe_2O_3	3.88	2.33	2.24	2.60	3.80
FeO	7.75	8.00	8.20	8.52	6.69
MgO	8.89	8.02	8.60	5.60	5.23
CaO	10.48	10.26	8.16	10.30	9.29
Na_2O	2.51	2.02	2.26	1.72	2.77
K_2O	0.46	0.35	0.57	0.15	0.75
H_2O +	n.d	—	0.18	0.87	0.16
H_2O −	n.d	—	0.22	0.42	0.34
TiO_2	0.73	0.90	1.18	0.73	0.72
P_2O_5	0.15	n.d.	0.16	0.06	0.03
MnO	0.54	0.25	0.20	0.16	0.16
Ign. loss	—	0.68	—	—	—
SrO	n.d.	n.d.	0.16	n.d.	0.14
Total	100.17	99.54	99.913*	100.60	99.78
		Norm			
Q	—	—	3.00	4.62	5.88
Or	2.78	2.22	3.34	1.11	4.45
Ab	20.96	16.77	18.86	14.68	23.58
An	33.08	38.36	31.69	44.23	30.58
Wo	7.66	5.22	3.60	2.91	6.61
En	11.20	18.70	21.50	13.96	13.10
Fs	5.28	11.09	11.62	12.66	8.18
Fo	7.70	0.98	—	—	—
Fa	4.28	0.61	—	—	—
Mt	5.57	3.25	3.25	3.70	5.57
Il	1.37	1.67	2.28	1.37	1.37
Ap	0.34	—	0.34	tr.	0.00

* Including BaO 0.039, ZrO_2 0.021, $(Ce, Y)_2O_3$ 0.003

1. Olivine basalt with a very little xenocrysts of quartz (HK 47041011) from the northwestern foot of Sukumo-yama. Analyst, analysts of the Government Chemical and Industrial Research Laboratory, Tokyo.
2. Olivine basalt (HK 47040702) from a quarry 1.5 km northwest of Zyōbosi crater. Analyst, T. KATSURA (personal communication).
3. Same as 2. Analyst, K. NAGASHIMA. Quoted from NAGASHIMA (1953).
4. Olivine-augite-hypersthene basalt from the western foot of a hill halfway between Itō and Ogi. Analyst, S. TANAKA. Quoted from TSUYA (1937).
5. Olivine-augite basalt with a little xenocrysts (HK 47040706) from a road-cut 1 km northwest of Ogi. Analyst, K. NAGASHIMA. Quoted from NAGASHIMA (1953).

Table 2. Continued.

Volcano	Komuro-yama				Ōmuro-yama	Daino-yama
No.	6	7	8	9	10	11
Type of assemblage	IIId	IIId	IIId	IIa→d	IIId	Xd
SiO_2	56.03	53.26	54.12	58.02	57.74	58.17
Al_2O_3	17.90	18.01	17.90	16.94	17.70	15.85
Fe_2O_3	1.69	2.57	2.28	2.12	1.95	4.38
FeO	6.16	6.38	6.53	5.08	5.32	4.35
MgO	4.33	5.58	5.61	4.31	5.00	4.63
CaO	8.40	8.26	8.24	6.86	9.14	8.04
Na_2O	2.83	2.55	2.78	3.40	1.59	2.64
K_2O	0.83	0.72	0.77	1.05	0.72	1.12
H_2O +	0.60	1.00	0.53	0.80	0.09	0.07
H_2O −	0.27	0.27	0.19	0.24	0.16	0.18
TiO_2	0.74	0.87	0.85	0.72	0.28	0.45
P_2O_5	0.22	0.23	0.22	0.19	0.04	0.16
MnO	0.11	0.18	0.20	0.17	0.09	0.18
Total	100.11	99.88	100.22	99.90	99.82	100.22
Norm						
Q	9.85	7.38	6.60	11.22	16.92	15.12
Or	5.01	3.89	4.45	6.67	3.89	6.67
Ab	24.12	21.48	23.58	28.82	13.10	22.53
An	33.66	35.58	33.92	27.52	39.48	28.08
Wo	2.55	1.86	2.55	2.32	2.44	4.64
En	10.74	14.00	14.00	10.80	12.50	11.60
Fs	8.97	8.58	8.98	6.86	7.66	3.96
Mt	2.55	3.71	3.25	3.02	3.02	6.50
Il	1.37	1.67	1.67	1.37	0.61	0.91
Ap	0.65	0.34	0.34	0.34	0.00	0.34

6. Olivine andesite with a little xenocrysts from the eastern foot of the cone of Komuro-yama. Analyst, S. TANAKA. Quoted from TSUYA (1937).
7. Olivine andesite with a little xenocrysts (HK 54011506) from a quarry just west of Kawana. Analyst, T. KATSURA (personal communication).
8. Same as 7, with a moderate amount of xenocrysts. Analyst, T. KATSURA (personal communication).
9. Same as 7, with a considerable amount of xenocrysts. Analyst, T. KATSURA (personal communication).
10. Olivine andesite with a little xenocrysts (HK 49011206a) from Akazawa, south of Yawatano. Analyst, K. NAGASHIMA. Quoted from NAGASHIMA (1953).
11. Augite andesite with a moderate amount of xenocrysts (HK 49010901a) from the western flank of Daino-yama. Analyst, K. NAGASHIMA. Quoted from NAGASHIMA (1953).

being characterized by high Al_2O_3, low alkalies, and low $MgO/FeO + Fe_2O_3$ ratio.

The rock forming the dike in the scoria has a character intermediate between those of the pigeonitic and hypersthenic rock series. It is sparcely porphyritic olivine basalt (type IIIa→c) with phenocrysts of olivine and a little plagioclase in a groundmass consisting of plagioclase, augite, pigeonite, magnetite, a little olivine, and interstitial cristobalite and brown glass. A few crystals of hypersthene are also found in the groundmass. The groundmass olivine is surrounded by reaction rim of clinopyroxene. Very sporadic xenocrysts of andesine may be seen in hand-specimens.

Basalts and andesites of the hypersthenic rock series. About 90% of the rocks of the Ōmuro-yama Volcano Group, namely, the rocks of the upper member and pyroclastic ejecta of the Umenoki-daira Volcano, the rocks of Zyōbosi, Ōmuro-yama, Komuro-yama, and Daino-yama, and the rock forming the flow in the Hatiga-kubo scoria all belong to this rock series. Scoria of Hatiga-kubo and Ōta are glassy olivine basalts possibly of the same series.

All the rocks are aphyric or sparcely porphyritic. Phenocrysts of olivine are nearly always present, while those of plagioclase may or may not occur. Phenocrysts of augite, hypersthene, and magnetite are found in some rocks in varying quantities.

The kinds of groundmass constituents are nearly uniform throughout basalts and andesites. They are plagioclase, augite, hypersthene, magnetite, ilmenite, and interstitial cristobalite and anorthoclase. Brown glass may occur in some rocks. Olivine is found in the most mafic basalt and also in glassy facies of basalts and andesites. It is always surrounded by reaction rim of pyroxene.

Olivine basalt (type III a→d) of Zyōbosi is the most mafic type, containing a little olivine in the groundmass together with cristobalite and glass. Two analyses made on the same specimen of this rock are given in Table 2, columns Nos. 2 and 3. The rock is just saturated with SiO_2. The discrepancy between the two analyses cannot be accounted for.

Augite-olivine andesite (type IVa→d) which forms the flow in the Hatiga-kubo scoria and olivine-augite basalt (type IVd) of Umenoki-daira represent a less mafic type. The former contains a little olivine and a considerable amount of glass in the groundmass. The analysis

of the latter given in Table 2, column No. 5, shows that the rock is transitional to andesite.

Olivine andesite (type IIId) of Ōmuro-yama is nearly identical in mineralogical and chemical compositions to that of Komuro-yama (type IIId). Some rocks contain considerable amount of glass. Analyses of the rocks of the two volcanoes are given in Table 2, columns Nos. 6, 7, 8, 9 and 10. Nos. 7, 8 and 9 are analyses of different parts of the specimen shown in Pl. III, Fig. 4. SiO_2 content increases with the quantity of xenocrysts of quartz and andesine contained in the rock. Therefore the difference between the three analyses are attributable to the difference in the amount of contamination of the salic plutonic rock. The vesicularity of the rock also increases with the quantity of the xenocrysts as is seen in Pl. III, Fig. 4.

Augite andesite of Daino-yama (type Xd) is holocrystalline and comparatively rich in xenocrysts. Its composition as shown in Table 2, column No. 11, is similar to that of the rock No. 9 which is also rich in xenocrysts.

Inclusions. Fragments of quartz diorite or granodiorite, ranging from 1 to 10 cm in diameter (Pl. III, Fig. 3), are found as inclusions in the Ōmuro-yama lava exposed at the northeastern margin of the village of Ike. The plutonic rock was originally made up of quartz and andesine, possibly with a small quantity of a mafic mineral and alkali-feldspar. However, as a result of reaction with the magma, the mafic mineral and alkali-feldspar have been entirely resorbed and andesine has been subjected to incipient melting along cleavages, and the inclusions now consist of isolated crystals of andesine with dust inclusions and quartz, together with abundant brown glass. In a more advanced stage of the reaction, the inclusions are changed to aggregates of irregular relics of quartz, recrystallized plates of tridymite, and irregular mosaic grains of anorthite which replace andesine, together with interstitial brown glass.

Scattered xenocrysts of quartz and andesine, most probably derived from the salic plutonic rock found as the inclusions, are universally present in the rocks of the hypersthenic rock series except in the Zyōbosi lava, although their quantity varies in different lavas and also in different parts of the same lava. Thus, in the Komuro-yama lava exposed in a quarry just northwest of Kawana, the xenocrysts are more concentrated in certain zones extending parallel to the plane

of the flow movement of the lava. The zones rich in xenocrysts probably represent partially resorbed inclusions of the plutonic rock drawn out by the flow movement. One of such zones is included in the middle of the specimen shown in Pl. III, Fig. 4. The analyses made on different parts of this specimen (Table 2, columns Nos. 7, 8, and 9) also reveal that the zone rich in xenocrysts (No. 9) has a composition corresponding to a mixture of a magma approximately

Table 3. Composition of parental magma of the Izu and Hakone province and average compositions of basalts of Ōsima Volcano.

No.	12	13	14
Rock type	Olivine basalt	Olivine basalt	Pyroxene basalt
SiO_2	48.73	52.52	52.59
Al_2O_3	16.53	15.14	15.05
Fe_2O_3	3.37	3.79	3.28
FeO	8.44	9.62	10.61
MgO	8.24	5.11	4.67
CaO	12.25	10.10	9.99
Na_2O	1.21	2.03	1.65
K_2O	0.23	0.43	0.40
TiO_2	0.63	0.83	1.41
P_2O_5	0.10	0.15	0.20
MnO	0.29	0.30	0.16
Total	100.02	100.03	100.01
	Norm		
Q	1.86	8.04	10.26
Or	1.11	2.22	2.22
Ab	9.96	17.29	14.15
An	39.20	30.86	32.53
Wo	8.70	7.66	6.84
En	20.60	12.80	11.70
Fs	12.14	13.60	14.52
Mt	4.87	5.57	4.87
Il	1.22	1.52	2.74
Ap	0.34	0.34	0.34

12. Parental magma of the pigeonitic rock series of the Izu and Hakone province. Average of two analyses of nearly aphyric olivine basalts. Quoted from KUNO (1953).

13. Average of three analyses of the pre-caldera lavas (or somma lavas) of Ō-sima. Quoted from KUNO (1953).

14. Average of six analyses of the post-caldera lavas (or central cone lavas) of Ō-sima. Quoted from KUNO (1953).

represented by the rock No. 7 (sparcely porphyritic with a little xenocrysts) and a certain salic pultonic rock.

The scattered xenocrysts of quartz are usually rounded in outline. Those of andesine are surrounded by broad zones rich in dust inclusions formed by incipient melting. Such zones often truncate the original zoning of the andesine crystals. Xenocrysts of hornblende are completely replaced by fine-grained aggregate of augite and hypersthene.

Although the xenocrysts are characteristic of the hypersthenic rock series, they are also found in a few rocks of the pigeonitic rock series, namely in the dike cutting through the Hatiga-kubo scoria and in the andesite lava of the lower member of the Umenoki-daira Volcano. In both cases, the groundmass contains hypersthene rimmed with clinopyroxene:—a feature transitional to that of the groundmass of the hypersthenic rock series.

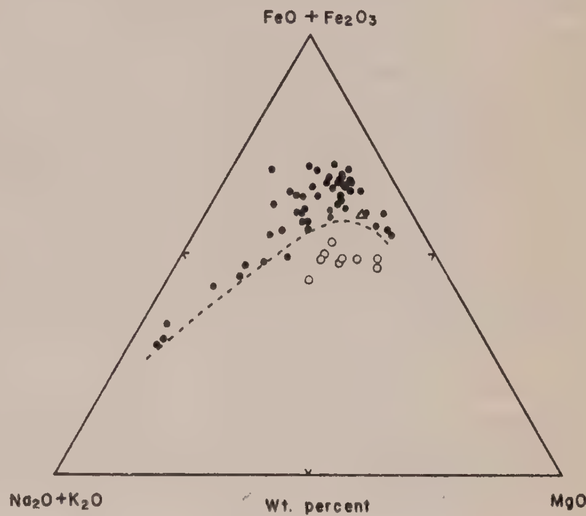

Fig. 6. Compositions of aphyric and nearly aphyric rocks of the pigeonitic rock series of the Izu and Hakone province (solid circles) and those of the hypersthenic rock series of the Ōmuro-yama Volcano Group (open circles). Triangle represents the porphyritic basalt of the former series of the Group.

Comparison with rocks of other volcanoes. In Table 3, columns Nos. 13 and 14, are given the average compositions of the somma lavas of the Ō-sima Volcano and that of the central cone lavas respectively. They are typical differentiates of the parental tholeiitic magma such as shown in column No. 12 of the same table (KUNO, 1953). According to the pyroxenic phase of the groundmass, they belong to the pigeonitic rock series. As the rocks are nearly aphyric, the analyses may be used for comparing the compositions of the magmas of the pigeonitic rock series with those of the hypersthenic represented by most of the rocks of the Ōmuro-

yama Group. The two are the nearest basaltic volcanoes erupted in the same general age.

The rocks of Ō-sima are lower in Al_2O_3, alkalies, and $MgO/FeO+Fe_2O_3$ ratio than those of Ōmuro-yama with the corresponding SiO_2 contents. Total iron is higher in the former than in the latter.

These differences exist generally between the rocks of the pigeonitic rock series and those of the hypersthenic. In order to show the ratio MgO : total iron : total alkalies, all available analyses of aphyric and nearly aphyric rocks of the pigeonitic rock series of the Izu and Hakone province, including those of Ō-sima, are plotted in Fig. 6 with solid circles. The rock No. 4 of Table 2, which is the only analysis of the rocks of the pigeonitic rock series of Ōmuro-yama, is plotted with a triangle. Although the rock is porphyritic, the point lies within the area of the pigeonitic rock series.

The rocks of the hypersthenic rock series of Ōmuro-yama are plotted in the same figure with open circles. They lie within an area quite distinct from the area of the pigeonitic rock series.

As has been discussed elsewhere (KUNO, 1953), the trend of variation of the compositions as shown by the solid circles in Fig. 6 is explained as due to fractional crystallization of the parental tholeiitic magma, while the trend shown by the open circles is interpreted as due to contamination of the magma by granitic rocks.

The rocks of the Ōmuro-yama Volcano Group furnish direct evidence of contamination in that the partially resorbed xenocrysts from salic plutonic rock are almost universally present in the rocks of the hypersthenic rock series but are absent in those of the pigeonitic.

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Plate III.

Explanation of Plate III.

Fig. 1. Sukumo-yama as seen from the northwest.

Fig. 2. Lava plateau of the Umenoki-daira Volcano as seen from the northwest. Komuro-yama in the background.

Fig. 3. A partially melted xenolith of quartz diorite or granodiorite in the Ōmuro-yama lava.

Fig. 4. Scattered xenocrysts of quartz and andesine in the Komuro-yama lava.

Fig. 1

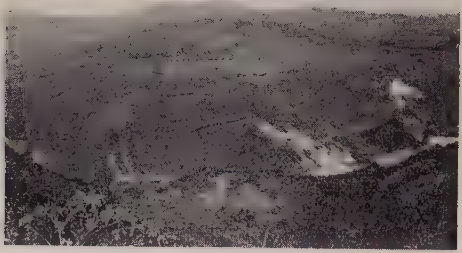


Fig. 2

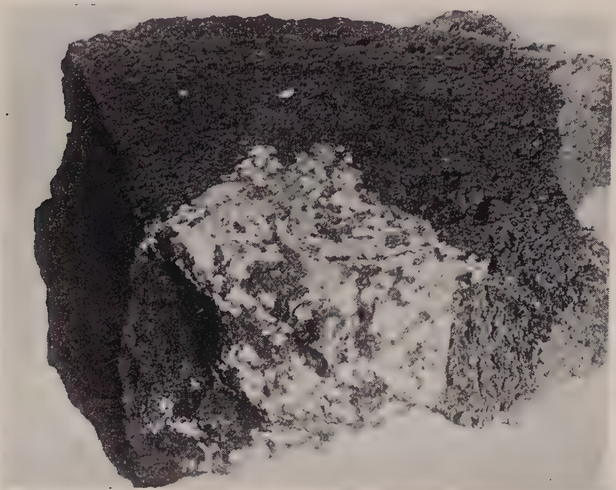


Fig. 3

Fig. 4

Unit Cell Dimensions of Synthetic Nepheline

By

Akiho MIYASHIRO and Tami MIYASHRIO

Natural nepheline may be regarded as a solid solution approximately of the following four molecules, of which the first two are predominant.

Sodium-nepheline = Ne = $Na_2Al_2Si_2O_8$
Kaliophilite = Kp = $K_2Al_2Si_2O_8$
Anorthite = An = $CaAl_2Si_2O_8$
Excess silica = Se = $Si_2Si_2O_8$

A hexagonal unit cell of nepheline contains 32 oxygen atoms.

In his celebrated study of natural nepheline and kaliophilite, BANNISTER(1) tried to correlate various physical constants of nepheline with its chemical composition. The relations are, however, so complicated that they could be established only partly. For example, he stated: "It has not been possible to trace any relationship between the lengths of the cell edges and the chemical composition, but if the cell volumes be considered there is a very small but definite increase with potassium content". In the present paper the writers

Table 1. Power Pattern of Synthetic Nepheline ($Ne_{81.7}$ $Kp_{18.3}$).

hk.l	Intensities	$2\theta°$ for $CuK\alpha_1$	d(Å)
11.0	5	17.6	5.03
20.0	20	20.4	4.35
00.2	60	21.1	4.20
20.1	75	23.0	3.87
12.0	50	27.12	3.285
20.2	100	29.53	3.022
30.0	40	30.85	2.896
12.2	30	34.67	2.585
22.0	10	35.77	2.5081
13.0, 22.1	10	37.35	2.405
20.3	40	38.26	2.3504
31.1	20	38.86	2.3155

Note: Reflections 22.0 and 20.3 are the most suitable for accurate measurement.

(1) F. A. BANNISTER: A chemical, optical and X-ray study of nepheline and kaliophilite. *Min. Mag. Vol. 22* (1931) pp. 569–608.

Table 2. Unit Cell Dimensions and Calculated Density of Synthetic Nephelines

No.		1	2	3	4	5	6	7
Mol. %	Ne	100.0	90.9	81.7	71.8	62.5	88.4	94.8
	Kp	—	9.1	18.3	28.2	37.5	—	—
	An	—	—	—	—	—	11.6	—
	Se	—	—	—	—	—	—	5.2
a (Å)		9.992	10.022	10.032	10.046	10.062	9.975	9.978
c (Å)		8.342	8.376	8.384	8.407	8.415	8.356	8.343
c/a		0.8349	0.8358	0.8357	0.8368	0.8362	0.8377	0.8361
V (Å^3)		721.3	728.6	730.7	734.8	737.9	720.1	719.3
ρ		2.615	2.616	2.635	2.649	2.665	2.613	2.601

Note: Cell volume $V=a^2c \sin 60°$. Density $\rho=1.66020 \sum M/V$.

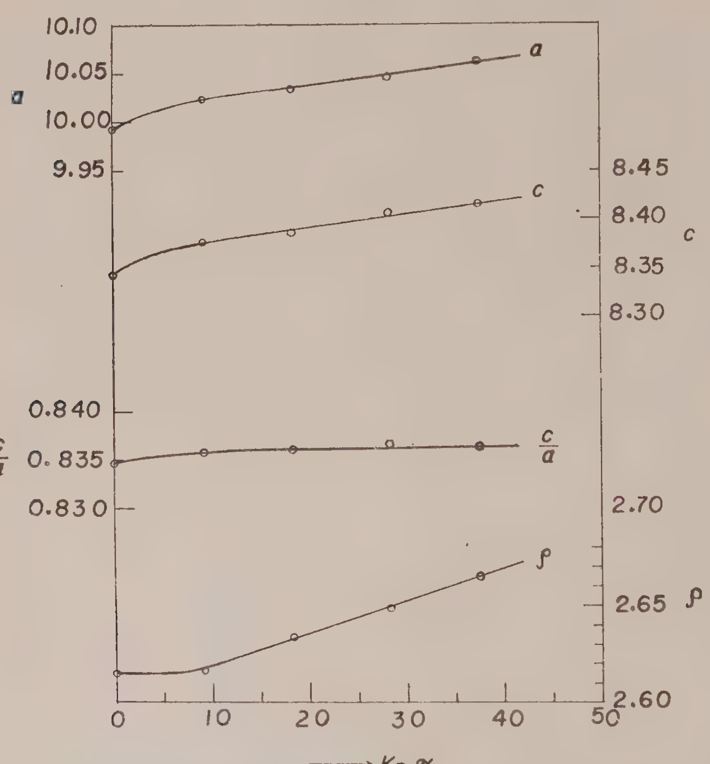


Fig. 1. Unit Cell Dimensions, Axial Ratio and Density of Nephelines Composed of Ne and Kp Molecules.

intend to establish the relations between the unit cell dimensions, axial ratio, and density on the one hand, and the chemical composition on the other, by using synthetic nephelines which have much simpler compositions than natural nepheline.

Solid Solution Series Ne—Kp

Sodic members of the solid solution series Ne—Kp were synthesized by sintering, around 1200°C for 20–40 hours, mixtures of chemicals Na_2CO_3, K_2CO_3, Al_2O_3, and SiO_2 in appropriate proportions. Powder spectrograms for CuKα radiation were taken by means of the Philips X-ray spectrometer.

The powder patterns are very similar to that of natural nepheline. The dimensions of the unit cells were calculated as shown in Nos.1–5 of Table 2, which also gives the axial ratios (c/a) and densities (ρ), calculated from the cell dimensions.

As is shown in Fig. 1, the unit cell and density become larger with the increasing potassium content. The axial ratio also tends to become larger slightly.

The accuracy in the measurement of cell edges is probably of the order of ± 0.010 Å.

Effects of Anorthite and Excess Silica Molecules

In order to elucidate the effects of the An and Se molecules which replace the Ne molecule in nepheline, the writers synthesized two compounds, as shown in Nos. 6–7 of Table 2, by the same procedure.

Comparison of Nos. 6–7 with No. 1 of Table 2, shows that the two molecules replacing Ne in nepheline decrease the edge a but increase the edge c, as shown in Fig. 2, which is constructed by assuming linear variations.

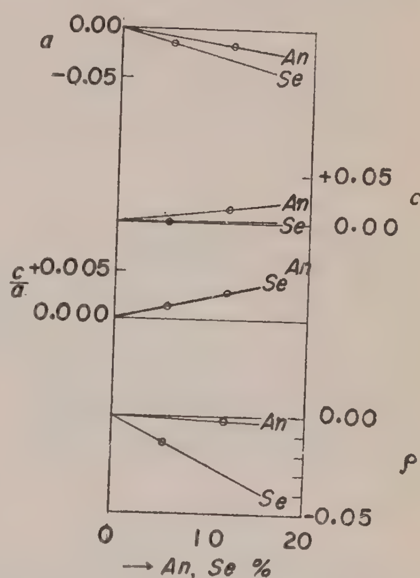


Fig. 2. Effects of An and Se Molecules Replacing Ne in Nepheline.

Table 3. Nepheline from Shinjozan

	Wt. %		
SiO_2	41.94	Mol. composition:	
Al_2O_3	34.26	Ne	71.4
Fe_2O_3	0.55	Kp	17.9
MgO	0.30	An	1.4
CaO	0.28	Se	9.3
Na_2O	15.82	Measured unit cell:	
K_2O	6.04		$a=10.014$ Å
H_2O_+	1.32		$c=8.392$ Å
H_2O_-	0.25		$c/a=0.8380$
Cl	0.09		
	100.85	Refractive indices	
Less O	−0.02		$\omega=1.541$
	100.83		$\varepsilon=1.537$

Note: Analysis by Nobufusa SAITO (*Jour. Chem. Soc. Japan, Vol. 64*, 1943, p. 1152).

Calculation of the Unit Cell Dimensions and Density of Natural Nepheline

The unit cell dimensions and density of any natural nepheline whose composition is known, can be calculated by using Figs. 1 and 2, if we may assume that the effects of these molecules are additive. As an example, Table 3 shows a chemical analysis and the molecular composition calculated therefrom, of a nepheline in nepheline-syenite pematite at Shinjozan in the Fukushinzan district, Korea. The unit cell dimensions may be calculated from the molecular composition as follows:

	a	c	c/a	ρ
Nepheline of composition $Ne_{82.1}$ $Kp_{17.9}$	10.035	8.389	0.8360	2.632
Effect of 1.4% An replacing Ne	−0.004	+0.002	+0.0003	−0.001
Effect of 9.3% Se replacing Ne	−0.026	+0.002	+0.0022	−0.024
Sum	10.005	8.385	0.8385	2.607

Thus, we obtain $a=10.005$ Å, $c=8.385$ Å, $c/a=0.8385$, and $\rho=2.607$. These values are in agreement, within the experimental error, with the directly measured values: $a=10.014$ A, $c=8.392$, and $c/a=0.8380$.

Acknowledgements

The writers' thanks are due to Drs. Hisashi KUNO and Hisao YAMADA and to Messrs. Shin'ichi IWAI and Kozo SUGIURA for their interest and help during this work.

Structural Control and Rock Alteration at the Nishiazuma Mine, Yamagata Pref., Japan

By

Hiromu MUKAIYAMA

Abstract

An appreciable amount of sulfur and iron-sulfide ores have been produced from platy formed replacement deposits of the Nishiazuma mine which are found in volcanic rocks of the Nishiazuma volcano. In the deposit, the control of sulfur mineralization is primarily structural. Selective replacement of tuff-breccia beds and presence of fractured zones which are thought to be the centre of mineralization are contributing factor. The high-grade ore bodies are usually formed in tuff-breccia beds where fractured zones cut them. Various kinds of altered rocks, such as sulfurized, pyritized, opalized, alunitized, and kaolinized rocks show zonal distribution around the high-grade ore bodies and the fractured zone.

Chemically, these altered rocks are characterized by a leaching process. A large proportion of Ca, Mg, Fe and alkali and a small proportion of Al and silica have been removed from original rock without change in the apparent volume of the rocks. Only two components, S and H_2O are found added in larger proportion from mineralizing fluid to altered rocks. Thermal gradients which are calculated by thermal data on andesite, show that differences of the temperature in each of the altered rocks, surrounding the centre of mineralization are not conspicuous. We can not consider therefore that differences of conditions on the formation of each of the altered rocks depend mainly on the differences in temperature.

Introduction

In the north-eastern part of Japan, a large number of sulfur and iron-sulfide deposits[1] are found in volcanic areas along the Nasu volcanic zone. These sulfur deposits may be classified into three types: that is, sublimation, sedimentary and replacement types and are formed by the action of volcanic gases and hydrothermal solutions, on and near the surface of volcanic rocks. The deposits of sublimation and sedimentary types are located in recent and older solfataric areas, and those of replacement type are emplaced in volcanic rocks near the surface.

Among these three types, the replacement type is economically the most important, because of its larger scale, and the greater part of

sulfur production of Japan comes from this type.

Sulfur deposits of the Nishiazuma mine are those of replacement type, and are found in andesitic rocks on the northern slope of Nishiazuma volcano at 1400 m. above sea-level. The mining camp is situated in Minamihara Mura, Minamiokitama Gun, Yamagata Pref., Japan.

The formation of high-grade ore bodies was controlled by the structure of country rocks, and various kinds of altered rocks which are thought to have been formed by the action of acidic solutions and gases are seen in mineralized areas. Results from the studies on the structural localization of the high-grade ore and rock alteration will be described in this paper.

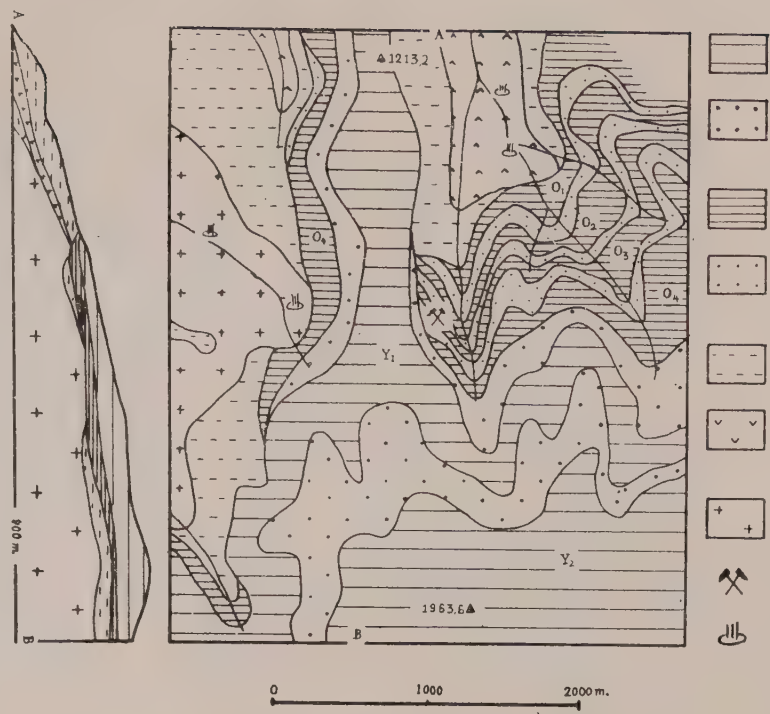
The writer is greatly indebted to Professors Takeo WATANABE, Toshio SUDO and Shuichi IWAO and Mr. Fukutaro HORI for their kind advice and guidance in the laboratory. Grateful acknowledgement is also made to Mr. Chiharu YAMAMOTO, the president of the Yamagata Kogyo Company and the staff of the company for their kind instruction in field work, and to Mr. ISHIDA, the chief of the analytical section of the Geological Survey of Japan for the analysis of the rocks and ores.

General geology

Effusive rocks of the Nishiazuma volcano, which are probably of late Tertiary to recent age, cover gneissose granites and Tertiary formation on the northern slope of the volcano. The volcanic rocks, distributed around the mine, consist of lava flows, tuff-breccia and tuff beds of augite-hypersthene andesite, and are divided into two groups, the one being the older group and the other the younger group.

The rocks of the older group consist of four lava flows with several beds of tuff-breccia and tuff, all of them being augite-hypersthene andesite. General trends of the flows and the beds run nearly NE–SW, and dips are to the N gently. Some parts of the rock were subjected to severe solfataric action and altered to various types of rock. Sulfur deposits are extensively developed in the altered rocks.

The rocks of the younger group consist of two lava flows of augite-hypersthene andesite and their pyroclastic equivalents. Members of the group elongate from E to W, dip to N and discordantly overlie the older group.



Younger group

1. Lava flow, Y_1, Y_2.
2. Tuff-breccia bed.

Older group

3. Lava flow, O_1, O_2, O_3 and O_4.
4. Tuff-breccia bed.

Tertiary formation

5. Tuff and tuff-breccia bed.
6. Rhyolite flow.

Granitic rock

7. Gneissose granite.
8. Ore deposit.
9. Hot spring.

m: meters.

Fig. 1. Geological map and section of the Nishiazuma mine.

The Tertiary formation is mainly composed of rhyolitic flows and tuffs (the so-called green tuff), and covers the granitic rocks unconformably. These rocks are generally disturbed, although the general trends of the formation run nearly E–W, with dips to the N.

Granitic rocks belong mainly to gneissose biotite granites, forming the base of this area.

Numerous fissures and quartz veinlets cut both the Tertiary rocks and granitic rocks, and some of the veins include a small amount of gold and silver. The fissures and the veins have E to W trend, and incline steeply to N or S. The age of fissuring and the formation of the veins might correspond to the period of igneous activities in middle Tertiary in northern Japan. The structural feature and distribution of the rocks are shown in Fig. 1.

Ore deposits

The ore deposits of this mine are a group of platy formed bodies of sulfur and iron sulfide ores in the older group of the Nishiazuma volcanic rocks. The country rocks of the deposit are severely affected by mineralizing solutions and gases originated from the Nishiazuma volcano, during the period of the activity. This resulted in the formation of the ores as well as of various kinds of altered rocks. The sulfur ore is a member of the altered rocks which is most strongly affected by sulfurization.

The ores are composed mainly of sulfur and opal, accompanied by a small amount of iron-sulfide minerals and alunite. The average content of sulfur in various parts of the ore deposits, varies from 27% to 32% and in the higher grade ores attains 35% or more.

Surrounding the high-grade ore body or the fractured zone, various kinds of altered rock, such as sulfurized, pyritized, opalized, alunitized and kaolinized rock, are zonally distributed.

Structural features of the deposit

Granitic rocks and tuff-breccias of Tertiary formation, which form the base of this area, have an undulate surface with a narrow valley elongated from E to W and inclined gently to W.

Three lava flows and pyroclastic rocks of the older group (the Honko, the Mogami and the Osawa lavas and their pyroclastic equivalents),

are observed in the valley. The Honko lava is in the lowest part and is covered by the Mogami lava and the uppermost Osawa lava. They are augite-hypersthene andesite, black in color and compact. A few crystals of quartz which are possibly regarded as xenocryst, are usually included in the lavas.

Two tuff-breccia beds which are inserted between these lavas are named the Mogami bed and the Osawa tuff-breccia bed, respectively. They consist mainly of angular breccias of augite-hypersthene andesite, and sometimes include small fragments of granitic rocks and rhyolitic rocks. In each of the two beds is inserted a bed of tuffaceous shale or sandstone, being brown in color and containing carbonaceous matter. Carbonized leafs of bamboo-grass and latifoliate trees are frequently found in the bed.

In all of these rocks are well developed three fractured zones, with a trend E to W. A large number of sulfur, iron-sulfide, opal,

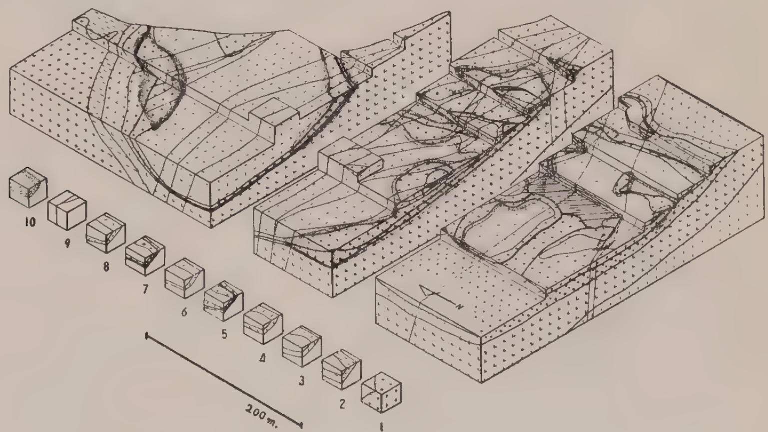


Fig. 2. The mode of occurrence of high-grade ore and structure of country rocks.

1. Granitic rock.
2. Tuff and Tuff-breccia bed of Tertiary formation.
3. Rhyolite flow of Tertiary formation.
4. Honko lava.
5. Mogami tuff-breccia bed and shale.
6. Mogami lava.
7. Osawa tuff-breccia bed and shale.
8. Osawa lava.
9. Fracture.
10. High-grade ores.

m: meters.

and clay veins and fissures are exclusively developed in those fractured zones.

Two principal structural environments favorable to the emplacement of sulfur and iron-sulfide ores are recognized. Introduction of sulfur and iron-sulfide took place along the fractured zones and they were deposited selectively in tuff-breccia between lava flows. Notable bodies of the high-grade sulfur ore are usually emplaced in those parts where the fractured zones cut the tuff-breccia beds. Thus the formation of the high-grade ore bodies were clearly controlled by the structure of country rocks, as is shown in Figs. 2 and 3.

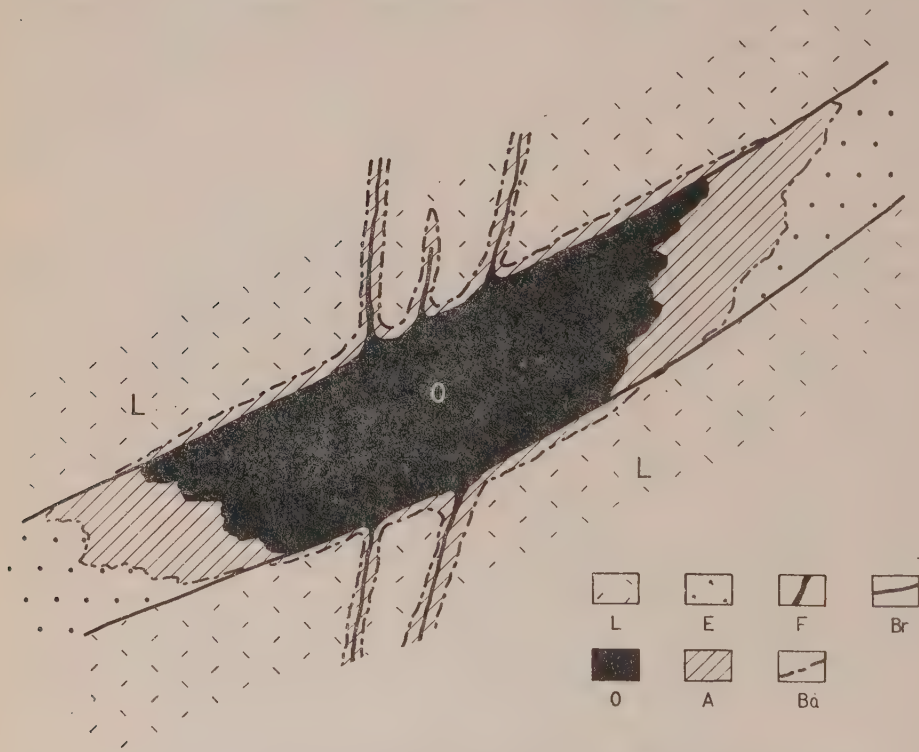


Fig. 3. Idealized section showing the structural localization of the high-grade ores and of altered rocks.
L: Lava flow E: Tuff-breccia bed O: Ore A: Altered rocks
F: Fracture Br: Rock boundary Ba: Boundary of altered zone

Rock alteration

Country rocks of the sulfur deposit have been strongly affected

by the action of mineralizing solutions and gases which issued during the period of activities of the Nishiazuma volcano. According to the characteristic mineral[2-11,16] in these altered rocks, they may be classified into (1) sulfurized rock (sulfur ore and iron-sulfide ore), (2) pyritized rock (iron-sulfide ore), (3) opalized rock, (4) alunitized rock and (5) kaolinized rock.

These altered rocks are zonally distributed around the high-grade ore bodies and fractures in and near the fractured zones. The most strongly sulfurized rock (1) may represent the centre of mineralization. Arround this centre, pyritized rock, opalized rock, alunitized rock and lastly kaolinized rock are zonally arranged.

Each altered rock is very rich in opaline silica and the characteristic mineral of alteration, according to the writer's field and petrographic observations.

No evidence that suggests a change in the apparent volume of rock in the processes of alteration have been recognized.

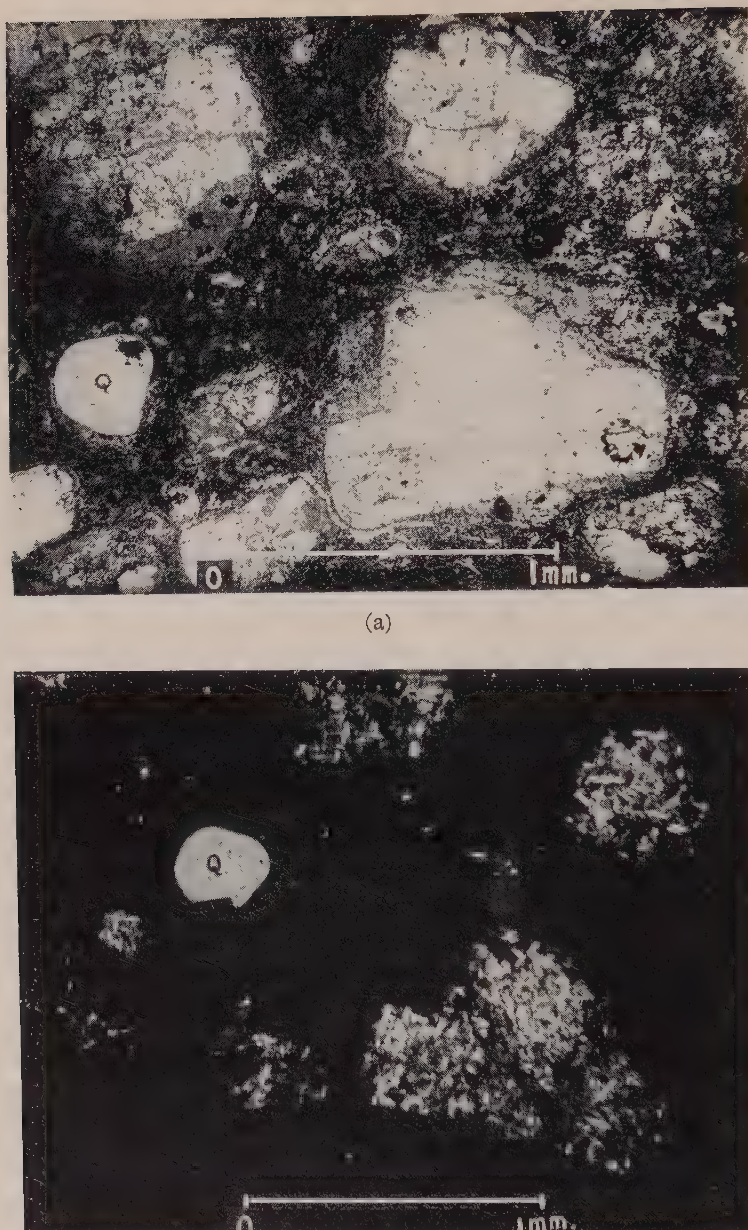
Microscopic observations of altered rocks

The texture and structure of the original andesite generally remain in all altered rocks except in the high-grade ore. For example, the porphyritic structure has been well preserved in the altered rocks and may be easily seen as in Fig. 4.

1. Sulfurized rock consists mainly of opal and sulfur, accompanied by iron sulfides (marcasite and pyrite) and alunite. In the high-grade ore (highly sulfurized rock) sulfur occurs as a large crystal replacing phenocrysts as well as the groundmass of the original rock, resulting in obliteration of the original texture and structure. In the low grade ore it replaces the phenocryst while the groundmass were selectively replaced by opal. Thus the original texture and structure are clearly seen.

2. Pyritized rock includes opal and iron-sulfides, together with subordinate quantities of sulfur and alunite. Small grains of iron-sulfide minerals are found sparsely in the highly opalized rock, especially in the groundmass. Iron-sulfides are also found in the relics of phenocryst of pyroxene. Small veinlets of iron-sulfides often cut the rock.

3. Opalized rock contains small amounts of iron-sulfides and alunite, preserves also its original texture and structure.



(a)

(b)

Fig. 4. Mode of occurrence of alunite. (a) // nicol, (b) +nicol. Q: Quartz.

4. Alunitized rock consists mainly of opal and alunite with small amounts of iron-sulfides and kaoline minerals. Alunite usually replaces phenocrysts and occurs usually as aggregates of small fibrous crystals. A small amount of alunite is also found sporadically in the opalized ground mass and veinlets of alunite with opal, of probably segregation origin are occasionally observed in the rock. Kaoline minerals may be the relict portion of the kaolinized rock*. Mode of occurrence of alunite is shown in Fig. 4 (a), (b).

5. Kaolinized rock is largely made up of opal and such kaoline minerals as kaolinite and halloysite. The rock rich in halloysite is usually developed in outer zone.

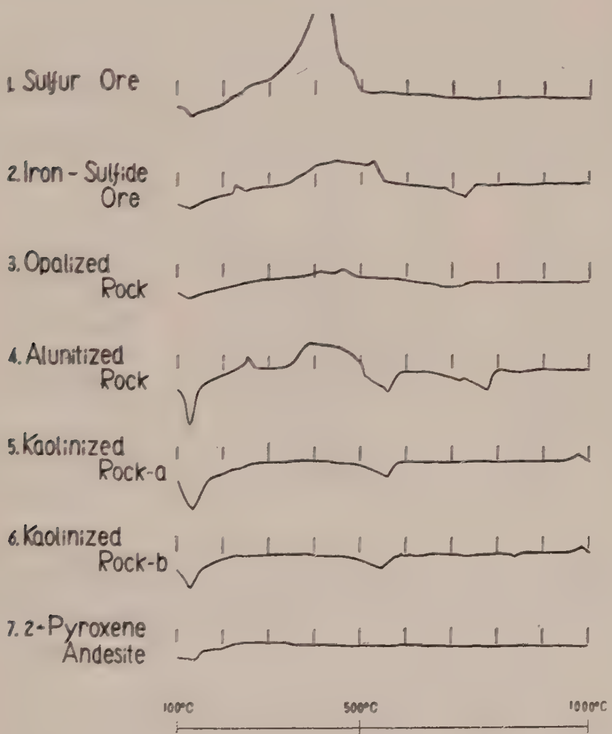


Fig. 5. Differential thermal curves of altered rocks.
(Number of the rock is the same as that shown in Fig. 6.)

* Alunitized rock occurring near the boundary between alunitized and kaolinized zones, includes blocks and small clots of kaolinized rock. These blocks and small clots are usually corroded due to alunitization.

Quartz in altered rocks

The original augite-hypersthene andesite usually contains xenocrysts of quartz. They have been preserved even in the most strongly altered rock. While the other constituents of the original andesite are all affected by mineralizing agents.

Studies of altered rocks by differential thermal analysis

In order to determine exactly the minerals which are included in altered rocks, D.T.A. experiments were carried out on each of the altered rocks. For example, results of the experiments on the altered rocks which are collected in Mogami lava are shown in Fig. 5. Mode of occurrences of the samples are shown in Fig. 6. As shown in Fig. 5 these altered rocks are characterized by the presence[12,13] of characteristic minerals in different altered rocks. These characteristic minerals are also observed under the microscope.

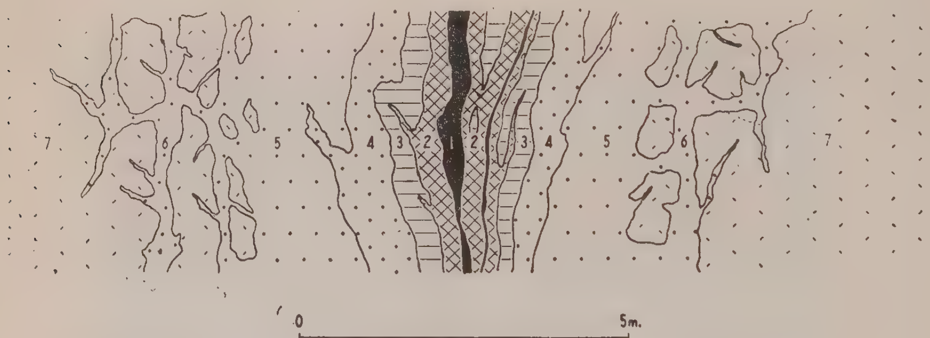


Fig. 6. Mode of occurrence of altered rocks in both sides of a fracture (Number of the rock is the same as that shown in Fig. 5.)

Chemical studies of altered rocks

Chemically, as is evident from Table 1, 2, 3 and Figs. 7, 8 the altered rocks are characterized by a leaching process with addition of S and H_2O. A large proportion of Ca, Mg, Fe and alkali and a small proportion of Al and silica are removed from original rocks, and S and H_2O are added to altered rocks from mineralizing fluid.

Samples for chemical studies were collected from a narrow altered zone, 3 m. in width, in order to insure the homogeneity of physical and chemical properties of original rocks. The altered zone is represented on both sides of the small fracture, traversing the Mogami

Table 1. Chemical composition, apparent specific gravity and true specific gravity of altered rocks.

Comp. (%)	Sulf. R.	Pyri. R.	Opal. R.	Alun. R.	Kaol. R.a	Kaol. R.b	Orig. R.
SiO_2	49.62	78.42	86.24	50.40	66.45	61.14	57.45
TiO_2	0.98	1.07	1.20	1.18	1.05	1.38	0.80
Al_2O_3	1.11	1.55	2.29	15.13	15.27	16.93	16.45
Fe_2O_3	1.19	0.81	0.67	3.71	2.30	2.00	2.54
FeO	0.54	0.35	0.14	0.79	0.25	0.46	5.69
Fe	—	2.54	0.13	2.04	—	—	—
MnO	nd.	nd.	nd.	—	—	—	0.06
CaO	0.04	0.08	0.03	0.02	0.17	0.09	7.13
MgO	0.03	0.07	0.07	0.09	0.15	0.16	4.30
BaO	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Na_2O	0.02	0.05	0.02	0.15	0.87	0.74	2.14
K_2O	0.03	0.03	0.02	0.28	2.12	2.12	1.24
$+H_2O$	3.70	4.30	3.65	7.31	5.22	5.79	0.78
$-H_2O$	4.83	5.41	4.42	9.42	4.78	6.81	1.02
SO_3	2.99	2.10	0.94	6.04	1.34	2.17	0.10
Free S	34.83	0.10	0.02	1.07	0.02	0.01	0.01
Comb. S	0.0	2.92	0.15	2.35	0.0	0.0	0.0
Total	99.92	99.81	100.0	99.99	100.0	99.81	99.72
True S.G.	2.1	2.08	2.13	2.40	2.38	2.35	2.76
Apparent S.G.	1.72	1.84	1.57	1.53	1.43	1.39	2.73
Porosity (%)	18.1	11.6	26.2	36.2	39.8	40.8	1.0

Table 2. Weight of each components in 1 cm.^{3} of altered rocks.

Comp. (gr.)	Sulf. R.	Pyri. R.	Opal. R.	Alun. R.	Kaol. R.a	Kaol. R.b	Orig. R.
SiO_2	0.854	1.443	1.354	0.771	0.950	0.850	1.568
Al	0.009	0.014	0.018	0.112	0.105	0.114	0.216
Fe	0.022	0.062	0.011	0.080	0.026	0.024	0.169
Ca	0.001	0.001	0.000	0.000	0.002	0.001	0.139
Mg	0.000	0.001	0.001	0.001	0.001	0.001	0.117
Alk.	0.001	0.001	0.001	0.005	0.034	0.033	0.071
TiO_2	0.017	0.020	0.019	0.018	0.015	0.019	0.022
Free S	0.599	0.002	0.003	0.016	0.000	0.000	0.000
Comb. S	0.000	0.054	0.002	0.036	0.000	0.000	0.000
$+H_2O$	0.064	0.079	0.057	0.112	0.075	0.081	0.028

Table 3. Differences of weight of main components between original and each altered rocks. (Calculated from Table 2).
− : subtraction, + : addition.

Comp. (gr.)	Sulf. R.	Pyri. R.	Opal. R.	Aluni. R.	Kaol. R.a	Kaol. R.b
SiO_2	−0.714	−0.125	−0.214	−0.797	−0.618	−0.718
Al	−0.207	−0.202	−0.198	−0.104	−0.111	−0.102
Fe	−0.147	−0.107	−0.158	−0.089	−0.143	−0.145
Ca	−0.138	−0.138	−0.139	−0.139	−0.137	−0.138
Mg	−0.117	−0.116	−0.116	−0.116	−0.116	−0.116
Alk.	−0.062	−0.070	−0.070	−0.066	−0.037	−0.038
Free S	+0.599	+0.002	+0.003	+0.016	0.000	0.000
Comb. S	0.000	+0.054	+0.002	+0.036	0.000	0.000
$+H_2O$	+0.036	+0.051	+0.029	+0.084	+0.047	+0.053

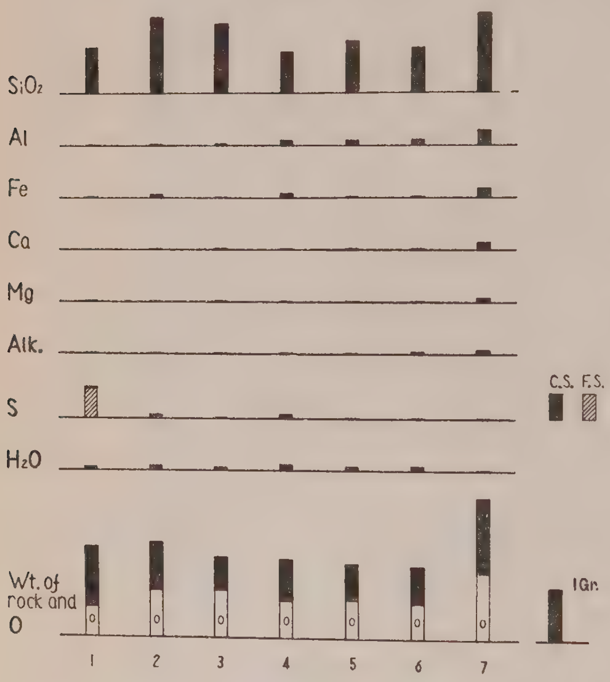


Fig. 7. Chemical composition of rocks expressed in grams per cubic centimeter.
1: Sulfurized rock 2: Pyritized rock
3: Opalized rock 4: Alunitized rock
5, 6: Kaolinized rock 7: Original rock
C.S.: Combined sulfur F.S.: Free sulfur
Number of each rock is the same as that shown in Fig. 6.

lava from which the samples are collected. Near the fracture, the Mogami lava is about 20 m. thick and lies between the two large bodies of high-grade ore which replaced the Mogami and Osawa tuff- breccia bed. The fracture lies between these two bodies and may represent a main pass of the mineralizing solutions and gases. Mode of occurrences of each of the altered rocks on both sides of the fracture are shown in Fig. 6.

Chemical composition, specific gravity, porosity and chemical composition of the original and altered rock expressed in

grams per cubic centimeter are shown in Tables 1, 2 and Fig. 7. Differences in this composition between each of the rocks and original rock are shown in Table 3 and Fig. 8.

From these values, the writer can draw the following conclusions that the alterated rocks were formed by the action of mineralizing solutions and gases without change in apparent volume of original rock and that the principal process of alteration was a leaching process.

The main components removed from the original rock by alteration, are Ca, Mg, Fe and alkali and a small proportion of Al and silica.

The main components which were added to the altered rocks from mineralizing solution were S and H_2O.

Silica, the main constituent of original rock was moved slightly, and consequently a larger proportion of it remained in each altered rock, except in sulfurized rock where a large proportion of it was removed.

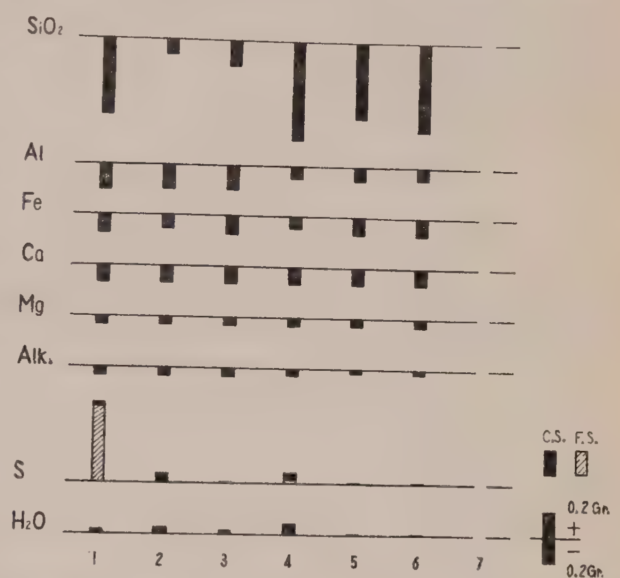


Fig. 8. Difference of the weight of each components between each of the rocks and original rock. (Number of the rock is the same as that shown in Fig. 7.)

Aluminum was considerably leached from sulfurized, pyritized, and opalized rocks, but in alunitized rock and kaolinized rock subtraction of Al was not marked in larger proportion.

Most of the iron was removed from all of the altered rocks, and calcium and magnesium were subtracted from the original rocks abundantly, and a small amount of them remained in altered rocks.

Alkali was also removed from the altered rocks.

Sulfur which was added to sulfurized rock from mineralizing solution and gases, formed native sulfur and became one of the main

consistuents of this rock. In the pyritized rock, it became mainly iron-sulfide minerals and in alunitized rock, alunite and iron-sulfides.

H_2O was added to all altered rocks, in especially large amounts to alunitized rock.

On the thermal gradient around the centre of mineralization

In order to consider the thermal relation during the alteration processes, the writer calculated the thermal gradients on both sides of the fracture which are shown in Fig. 6 by the following equation and thermal data.

$$\theta=\theta_0+(\theta'-\theta_0)\frac{2}{\sqrt{\pi}}\int_0^{\frac{2}{2\kappa\sqrt{t}}} e^{-\beta^2}d\beta \qquad \begin{array}{l}(\theta)_{x=0}=\theta_0 \\ (\theta)_{t=0}=\theta'\end{array}$$

κ^2 : The thermal diffusivity of andesite -0.012[14].

The temperature of the mineralizing fluid is presumed to be constant during the processes of mineralization. The writer also supposes that the mineralization processes through this fracture continued for many years[16], at least 50 years.

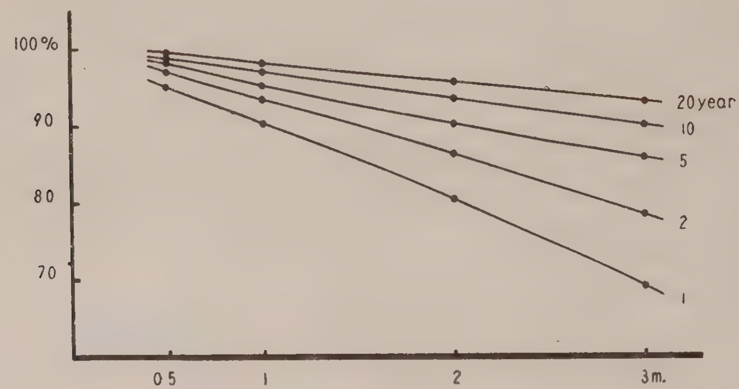


Fig. 9. Calculated thermal gradients on both sides of the fracture.

From the results of calculations shown in Fig. 9, it may be concluded that the country rocks of the deposit would be heated rapidly and temperature gradients are also enfeebled more swiftly, especially in the altered rocks where solutions and gases are migrated and thus the temperature gradient is not an important factor in the formation

of different types of altered rocks. Chemical natures of the rocks and the mineralizing fluid are the most important factors.

Conclusion

The ore deposits of the Nishiazuma mine are platy in form and composed mainly of native sulfur and opaline silica which replaced the andesitic lava flows and tuff-breccia beds of the Nishiazuma volcano.

The formation of high-grade ore bodies was generally dependent on the nature and structure of original rocks and on the distribution of the fractured zones along which the mineralizing fluids uprose. High-grade ore bodies have been usually localized along or near the fractured zone, especially in tuff-breccia beds.

Surrounding the high-grade ore or the fractured zone, various kinds of altered rock, such as sulfurized, pyritized, opalized, alunitized and kaolinized rocks are zonally distributed. Sulfur ore is also a member of these altered rocks and it has been affected by remarkable sulfurization.

The altered rocks keep the texture and structure of original rock very well, and no evidence that shows a change in apparent volume of the original rocks during the processes of mineralization has been found. The altered rocks are generally more porous than the original rock. Therefore their porosities must have been changed.

As to the behavior of each component in altered rock during the mineralization, it is shown that the greater part of silica, the main consistuent of the original rock, was not removed in a large proportion during the processes of mineralization. The main components, which are removed from the original rock, are Ca, Mg, Fe, alkali and a small proportion of Al. Subtraction of Al and silica was progressively stronger from the outer zone toward the centre of mineralization. The main components added to the altered rocks, are S and H_2O.

Thermal gradient of andesite calculated by the writer shows that differences of conditions in the formation of each rock were not caused by the differences in temperature.

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Stratigraphical Boundary Between the Pliocene and Pleistocene Strata in the Bôsô Peninsula, South Kantô, Japan

By

Sunao OGOSE

I. Introduction

The Bôsô Peninsula, South Kantô, Japan, is a classical locality of the so-called "Musasino formation" named by M. YOKOYAMA[29] in 1920. The formation is extensively distributed and most typically developed within the central and northern parts of the Peninsula.

A great number of fossils of molluscs and foraminifers, together with some of corals, echinoids, brachiopods, bryozoans, crustaceans and land mammals, etc. are found in various horizons of the formation. These fossils have already been studied in detail by many authors from various viewpoints such as of paleontology, biochronology, paleoecology, paleoclimatology and sedimentology. Based on these fossil evidences, the formation is generally regarded at present as belonging to the Japanese Pliocene and Pleistocene in age.

Thus, our knowledge concerning the fossil contents of the formation is fairly complete, but still much remains to be done on the studies of meaning of these fossils as an indicator of geological age of strata.

For this reason, the boundary between the Pliocene and Pleistocene strata in the Bôsô Peninsula is extremely difficult to draw.

II. Different Opinions Concerning the Stratigraphical Boundary Between the Pliocene and Pleistocene Strata in the Bôsô Peninsula

Separation of the Pleistocene strata from the Pliocene in the Bôsô Peninsula has hitherto been carried out by many authors who have engaged in geological and paleontological studies of the Peninsula. Up to the present, different opinions concerning the separation have already been published by various authors as shown in Table I.

However, the criteria controlling such separation have been insufficiently explained.

III. Stratigraphy of the District of Ônuki-mati, Sanuki-mati and Minato-mati, Kimitu-gun, Tiba Prefecture

Until quite recently, our knowledge concerning the stratigraphy of the district of Ônuki-mati, Sanuki-mati and Minato-mati (see Fig. 1) where the Pliocene and Pleistocene strata develop together has

Fig. 1

considerably been confused, as shown in Table II. This confusion has inevitably brought to our scientific world different opinions concerning the stratigraphical boundary between the Pliocene and Pleistocene strata in the Bôsô Peninsula.

As a result of the recent studies by the writer and A. HUZIWARA[1], however, the stratigraphic succession of the above-mentioned district is fairly clarified as shown in Table III.

Among these strata tabulated below, the Miura group is the oldest in age. Although this group is extensively distributed in the Miura Peninsula as well as in the middle and southern parts of the

Table III. Stratigraphic succession of the district of Ônuki-mati, Sanuki-mati and Minato-mati, Kimitu-gun, Tiba Prefecture, South Kantô, Japan

(A. HUZIWARA & S. OGOSE, 1952)

<table>
<tr><th colspan="2">Age</th><th colspan="3">Stratigraphic succession (in descending order)</th></tr>
<tr><td>Holocene</td><td>K</td><td colspan="3">Minato fossil-bearing silt (Raised beach deposit)</td></tr>
<tr><td>Upper Pleistocene</td><td>J_3</td><td colspan="3">Kantô volcanic ash (So-called Kantô loam) and Terrace sand and gravel</td></tr>
<tr><td rowspan="2">Middle Pleistocene</td><td>J_{1b}</td><td rowspan="5">Narita group</td><td rowspan="2">Middle division</td><td>Hitomi alternation of sand and mud</td></tr>
<tr><td>J_{1a}</td><td>Mihuneyama sand</td></tr>
<tr><td rowspan="3">Lower Pleistocene</td><td rowspan="3">I_2</td><td rowspan="3">Lower division</td><td>Sunami alternation of sand and silt</td></tr>
<tr><td>Sanuki silt</td></tr>
<tr><td>Nagahama sand and gravel</td></tr>
<tr><td rowspan="5">Upper Pliocene</td><td>I_1</td><td rowspan="5">Miura group</td><td rowspan="3">Akimoto subgroup</td><td>Itiziku sand / Nakazeki fine sand</td></tr>
<tr><td rowspan="4">H_2</td><td>Iwasaka fine sand</td></tr>
<tr><td>Komaba sand and gravel</td></tr>
<tr><td rowspan="2">Seki subgroup</td><td>Tômiya tuffaceous sandstone</td></tr>
<tr><td>Takeoka tuff-breccia</td></tr>
</table>

Bôsô Peninsula, only its upper division, i.e., the Akimoto subgroup, and middle division, i.e., the Seki subgroup, are developed in this district. As to the stratigraphic relation between the Itiziku sand and the Iwasaka fine sand, different opinions have hitherto been expressed by various authors. According to J. MAKIYAMA's opinion[8)~10)], the relation under consideration is unconformable, and the base of the Itiziku sand may be coincident with the beginning of the Pleistocene. On the other hand, the others[11), 18)~19)] have maintained that no decided stratigraphic break is recognized at the base of the Itiziku sand, and the two formations are completely conformable in stratigraphic relation. As a result of the writer's recent studies, however, it is quite clear that the relation in question is undoubtedly conformable and the lower part of the Itiziku sand may be considered to be a contemporaneous deposit with the upper part of the Iwasaka fine sand.

The Narita group which overlies the Miura group with a partial unconformity is well developed in the northern part of the district, and can be divided lithologically into five formations. Although the uppermost division of the Narita group is not developed in the district, there is no doubt that the Hitomi alternation of sand and mud, the uppermost division of the group in the district, is older than the so-called "Narita formation", i.e., the uppermost division of the group, which is extensively developed in the northern part of Tiba Prefecture. Therefore, the five formations ranging from the Nagahama sand and gravel to the Hitomi alternation of sand and mud may be considered to belong to the lower and middle divisions of the Narita group.

In the western part of the Peninsula, the Nagahama sand and gravel, i.e., the basal or marginal part of the Narita group, overlies the underlying Iwasaka fine sand and the Itiziku sand with a marked erosion-unconformity, while in the central part, the Mandano sand and gravel which is the eastern extension of the Nagahama sand and gravel overlies the underlying Itiziku sand and its eastern extensions, i.e., the Kosikiya alternation of sand and mud, etc., with no stratigraphic hiatus.

Concerning the detailed descriptions of these formations, the reader can refer to the writer's previous papers[1)~2)] and "Lexicon of Stratigraphic Names of Japan (Cenozoic Erathem)" recently published.

IV. Criteria Controlling Separation of the Pleistocene Strata from the Pliocene in the Bôsô Peninsula

Drawing of boundary between the Pliocene and Pleistocene strata in the Bôsô Peninsula has hitherto been made based on the criteria shown below:

1) The Lyellian method
2) The evolution of elephants
3) The difference of the contents of molluscan fossils
4) The climatic changes
5) The crustal movements

1) The Lyellian method

Based on the Lyellian method, M. YOKOYAMA[29)] concluded that the Upper Musasino formation is most likely not younger than Late

Pliocene in age.

Since then, J. MAKIYAMA[7] expressed an important opinion concerning the use of the Lyellian method. His opinion is as follows:

> "The Lyellian method may be applicable only in Europe. The percentage of recent species recognized in the Miocene of the tropical Pacific regions is said to be larger than that of the Lyellian Pliocene. In North Pacific region including Japan, this percentage is thought to be much reduced, for the change of ocean-currents since Pliocene time in this region have had significant effects on the fauna. Moreover, discrimination between species is much finer in modern conchology than in the day of DESHAYES".

As a result of the recent taxonomic studies in Japan, a great number of genera, subgenera and species of molluscan fossils reported hitherto from the Musasino formation have reexamined in definition and their scientific names have been largely revised to conform to the latest knowledge of taxonomic studies and to the rules of international nomenclature. Moreover, some species formerly believed to be extinct by M. YOKOYAMA[29)~30] are now regarded as living.

Although the marine molluscan fossils have been most extensively and sufficiently studied among various kinds of fossils from the Musasino formation in the Bôsô Peninsula, there is no doubt, for the above-mentioned reasons, that the Lyellian method is not so serviceable for the precise determination of age of the different strata belonging to the Musasino formation in the Bôsô Peninsula.

2) The evolution of elephants

J. MAKIYAMA[7] mentioned that the evolution of the Japanese elephants may be correspondent to that of the European. According to him, *Stegodon clifti* and *Parastegodon auroae* are undoubtedly of Pliocene in age while *Stegodon orientalis*, *Elephas namadicus naumanni* and *Elephas trogontherii* are most probably of Early Pleistocene.

Based on these fossil evidences, he concluded that the Lower Musasino is clearly Pliocene in age while the Upper Musasino certainly ranges from Late Pliocene to Pleistocene, as already mentioned by S. TOKUNAGA[24] and H. YABE[28]. And then, he presented a proposal that the line of demarcation between the Pliocene and Pleistocene strata in the Bôsô Peninsula is probably at the base of the Semata beds or at the top of the Kimitu beds in his classification of the Musasino series (=the Musasino formation of M. YOKAYAMA).

After that, Y. OTUKA[15] concluded that strata containing *Elephas*

namadicus naumanni are undoubtedly Pleistocene in age while the Akimoto group in which *Parelephas proximus*, an archetypal mammoth, is found may safely be considered to be Pliocene.

Although MAKIYAMA's opinion mentioned above is now accepted by almost all Japanese geologists and paleontologists, all of the fossil elephants up to now reported from various horizons of several localities in the Bôsô Peninsula are obtained without exception from the marine strata. In other words, these fossil elephants should be treated as more or less allochtonous ones*. Moreover, the precise localities and the modes of occurrence of these fossil elephants as well as the contents of associated marine invertebrate fossils are not always made clear because in most cases these fossil elephants were obtained by amateur geologists or natives instead of authoritative geologists.

Unfortunately, furthermore, the fossil elephants from the Bôsô Peninsula are too little known to make any valuable correlation, as shown in Table IV.

Therefore, in this case, these fossil elephants are not always appropriate for the precise determination of age of strata.

Table IV. List of the Fossil Elephants from the Bôsô Peninsula, Tiba Prefecture

1) *Parelephas proximus* (MATSUMOTO)
 Zizôdô-sita, Uehata, Akimoto-mura, Kimitu-gun. .... Hi
 Sankawa-dani, Ôtomi, Matuoka-mura, Kimitu-gun. .... 〃
 Sekinoya, Sekitoyo-mura, Kimitu-gun. .... 〃
 Higasi-higasa, Misima-mura, Kimitu-gun.......................... 〃
 Isone-misaki, Kokubo, Ônuki-mati, Kimitu-gun. .... ?
2) *Parelephas* cfr. *proximus* (MATSUMOTO)
 Yamawaki, Tamaki-mura, Kimitu-gun. .... Hi
 Kaburai-gawa, Oikawa-mura, Isumi-gun. .... Yu
3) *Stegodon sinensis* OWEN
 Nisiyatu, Naka, Ônuki-mati, Kimitu-gun. .... Sa
4) *Stegodon orientalis* OWEN
 Nagahama, Minato-mati, Kimitu-gun................................ Na
 Tôgane city. .... To

* It is notable that the mammalian fossils including *Elephas namadicus* and *Stegodon orientalis* found in the Nagahama sand and gravel, i.e., the basal or marginal part of the Narita group, are presumed to have been largely derived from the underlying Itiziku sand and are distinct climatologically from the associated molluscan fossils which indicate a cold current temperature.

5) *Elephas namadicus* FALC. & CAUT.
Nagahama, Minato-mati, Kimitu-gun. .... Na
Nakao, Kiyokawa-mura, Kimitu-gun. .... Oo
6) *Parelephas trogontherii* (POHLIG)
Nagahama, Minato-mati, Kimitu-gun. .... Na
Sanuki, Sanuki-mati, Kimitu-gun. .... Sa

Abbreviations:—

Hi .... Higasi-higasa mud, sand and gravel	Akimoto subgroup
Yu .... Yûgi sand	
Na .... Nagahama sand and gravel	Narita group
Sa .... Sanuki silt	
To .... Tôgane silt	
Oo .... Ôdorii sand and clay (Kiyokawa formation)	

3) The difference of the contents of molluscan fossils

H. UEDA[25)~26)] once expressed an opinion relating to the boundary between the Pliocene and Pleistocene strata in the Bôsô and Miura Peninsulas. According to him, the line of demarcation between the Pliocene and Pleistocene strata in the Bôsô Peninsula may be drawn at the base of the Sanuki formation, i.e., the lower division of the Narita group, or at the top of the Kakinokidai formation, i.e., the upper division of the Umegase group.

Based on UEDA's stratigraphical contributions, S. NOMURA[13)] stated that a remarkable difference of the contents of molluscan fossils is clearly recognized between the Narita and Umegase groups, that is to say, the fossil contents of the latter seem to be closely related to those of the Upper Pliocene strata in West Coast of North America, while the fossil contents of the former are composed mainly of the species characteristic to Japan.

It should be noticed here, however, that at present some defects are found in the contributions of H. UEDA and S. NOMURA on the stratigraphical side. As a result of the writer's recent studies, it is quite clear that much younger strata are included in UEDA's Umegase and Kurotaki groups*.

According to the writer's recent studies, the contents of molluscan fossils change very gradually from the upper part of the Akimoto subgroup, which is nearly identified with the Umegase group of

* To take a few examples, a part of the Kiwada formation, i.e., the middle division of the Kurotaki group, may be included in the Sanuki formation and, furthermore, the Tôgane silt of H. UEDA[26)] from which *Stegodon orientalis* was obtained is now regarded conclusively as belonging to the eastern extension of the upper part of the Sanuki formation (=the writer's Sunami alternation of sand and silt), but not of the eastern extension of the Kakinokidai formation.

H. UEDA[25)~26)], up to the lower part of the Narita group. There is no such a sudden change of the contents of molluscan fossils between the upper part of the Akimoto subgroup and the lower part of the Narita group as S. NOMURA[13)] believed to exist. Differences slightly noticeable of the contents of molluscan fossils between the upper part of the Akimoto subgroup and the lower part of the Narita group may be due merely to those of ecological or sedimentological conditions instead of biochronological factors.

Therefore, in this case, the contents of molluscan fossils are not always profitable for the precise determination of age of the Musasino formation.

4) The climatic changes

The climatic changes since the Pliocene Epoch are one of the most important factors in the separation of Pleistocene from Pliocene in Europe. The first appearance of the Pleistocene glacier will make a definite demarcation to Pliocene. In Japan, the mountains in the so-called "Japanese Alps" in Central Honsyû (the Main Island) and the Hidaka range in Central Hokkaidô were just high enough for the development of glaciers. Therefore, the criterion based on the climatic changes may be precisely applicable for the limnic or terrigenous strata formed in the above-mentioned mountains and their surroundings.

On the other hand, the climatic changes can not be a good criterion for the determination of age of marine strata such as the Narita group and the Akimoto subgroup in the Bôsô Peninsula because the relation between the changes of current temperature and those of climate on land is explained not sufficiently in the present state of our knowledge.

5) The crustal movements

Unconformity is generally regarded as an useful criterion to make a chronological boundary of older strata, but in the case under consideration, it seems not so valuable because no great regional crustal movement during Pliocene and Pleistocene Epochs is known in South Kantô including the Bôsô and Miura Peninsulas.

In the Miura Peninsula, there is a remarkable break between the Narita group (=the Yokohama group) and the underlying Miura group. Although the upper division of the Narita group has safely been assigned to Pleistocene in age by almost all recent Japanese geologists and paleontologists, there still remains a divergence of opinions on the boundary between the Pliocene and Pleistocene strata. According

to some, the above-mentioned unconformity may be somewhat adjacent to the boundary in question, if not coincident completely with it, while the others have maintained that this unconformity is slightly older than the close of Pliocene.

In the Bôsô Peninsula, on the other hand, no remarkable stratigraphical break has hitherto been ascertained throughout the strata belonging to the Narita group and the middle and upper divisions of the Miura group. The unconformable relationship between the two groups is only seen in the western part of the Peninsula. Accordingly, the boundary in question in the Bôsô Peninsula is quite difficult to recognize based on the unconformity.

Although not lending itself to an exact application or analysis, difference of geologic structure is also used as one of the criteria in making a chronological boundary.

In the Bôsô Peninsula, the Miura group is, as a rule, more complicated in structure than the overlying Narita group. Although the greater part of the Narita group is almost horizontal or dips gently to the north, some foldings and faultings are found in places especially in the environs of Sanuki-mati where the strata belonging to the lower part of the Narita group are locally inclined with more or less steep angles. On the other hand, the Akimoto subgroup is in some places also gently inclined to the north, likewise the greater part of the Narita group.

Therefore, the presence of foldings and faultings together with the steep inclination of strata provides no absolute criterion for the determination of their age.

V. The Writer's Opinion Concerning the Considered Boundary

As it is clear in the above-mentioned statements, one can find many difficulties in drawing the boundary between the Pliocene and Pleistocene strata in the Bôsô Peninsula.

In the present state of our knowledge, it can be only said that the lower division of the Narita group in which *Stegodon orientalis*, *Elephas namadicus*, *Parelephas trogontherii* have been found, is most probably Pleistocene in age, while the lower division of the Akimoto subgroup from which many specimens of *Parelephas proximus* have been obtained, may safely be considered to be Pliocene. As to

the age of the middle and upper divisions of the Akimoto subgroup, the writer can not determine it exactly in any way*.

At any rate, it is quite difficult to exactly determine the boundary between the Pliocene and Pleistocene strata in the Bôsô Peninsula.

According to the writer's present opinion, it is impossible, as a rule, to correlate the Japanese Pliocene and Pleistocene strata accurately with the European standard in unit of Epoch (Series) or Age (Stage) because the following reasons should be recognized.

1) Even in Europe, where the standard subdivision of Tertiary and Quaternary was at first established, still much remains to be studied on the boundary between the two Periods, and opinions are still conflicting with one another on this problem.

2) In the comparison with the Epochs belonging to Tertiary and older Periods, Pleistocene Epoch is believed generally to be very much shorter in an absolute time range. It may be said, therefore, that Pleistocene represents only the latest age of Pliocene.

Discussion of the stratigraphical boundary between Pliocene and Pleistocene, for that reason, should be restricted only in Europe and its neighbourhood in the present state of our knowledge.

Scientists who are to engage in geological and paleontological studies of the Japanese Younger Cenozoic strata must make at first correlation of the strata in sedimentary basins that are located separately in Japan. And then, it needs, for the present, to make a standard time-stratigraphic division of the Japanese Cenozoic strata.**

There remains one more question, however, that in what unit the time-stratigraphic correlations of the type sections of different sedimentary basins separately located with one another are possible. To prove exactly contemporaneity of strata located separately may be impossible at the present time. In fact, a divergence of opinions occurs concerning the correlation of the Miura and Narita groups in the western part of the Bôsô Peninsula with those in the Miura Peninsula, even though the two Peninsulas are closely located in geographical position with each other.

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* If *Stegodon orientalis*, *Elephas namadicus*, etc. obtained from the Nagahama sand and gravel were really derived from the underlying Itiziku sand, a part of the Itiziku sand may also be Pleistocene in age.

** From this point of view, the Letter Nomination proposed by N. IKEBE[5] is believed to be very much noteworthy for the biochronological studies of the Japanese Cenozoic strata.

VI. Conclusions and Acknowledgements

Although a great number of marine invertebrate fossils and some of land mammals have been found in various horizons of the Pliocene and Pleistocene strata in the Bôsô Peninsula, it is extremely difficult to determine where the boundary between the Pliocene and Pleistocene strata is.

The Nagahama sand and gravel, i.e., the basal or marginal part of the Narita group is probably Pleistocene in age, while the Higasi-higasa mud, sand and gravel which is a part of the basal or marginal beds of the Akimoto subgroup, may safely be considered to be Pliocene. *Stegodon orientalis* and *Elephas namadicus* have been obtained from the former while *Parelephas proximus*, an archetypal mammoth, from the latter. The writer can not determine exactly the age of the middle and upper divisions of the Akimoto subgroup in any way.

It is impossible to correlate the Japanese Younger Cenozoic strata accurately with the European standard in unit of Epoch (Series) or Age (Stage). It should be noted, finally, that discussions concerning the drawing of boundary between the Pliocene and Pleistocene strata in each sedimentary basin seem to be meaningless in the present state of our knowledge. International correlations must be done in future under the world-wide discussions.

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Table I. Different opinions concerning the stratigraphical boundary between the Pliocene and Pleistocene strata in the Bôsô Peninsula, South Kantô, Japan.

}: the boundary in question

Abbreviations:—ser.=series; st.=stage; subst.=substage z.=zone; b.=beds; g.=group; f.=formation

J. Makiyama (1929)

Akatuti f.		
Musasino ser.	Upper Musasino	Manzaki b.
		Tabata clay
		Semata b.
		Kimitu b.
	Lower Musasino	Naganuma b.
		Kosiba b.

J. Makiyama (1930a)

Akatuti f.	
Narita b.	Kiorosi st.
	Echinarachnius z.
	Manzaki st.
Semata b.	*Erodona* clay
	Kami-iwahasi st.
	Semata st.
	Kami-izumi st.
Kanôzan b.	Hitomi st.
	Itiziku st.
Sasage b.	
Miura b.	

J. Makiyama (1930b)

Akatuti f.		
Narita b.		Kiorosi st.
		Echinarachnius z.
		Manzaki st.
Semata b.		*Erodona* clay
		Kami-iwahasi st.
		Azu st.
		Kami-izumi st.
Kimitu ser.	Kanô-zan b.	Hitomi st.
		Itiziku st.
	Sasage b.	
Miura ser.	Yosino b.	
	Nokogiriyama b.	

J. Makiyama (1931)

Tatikawan st.	
Naritian st.	Kiorosi subst.
	Echinarachnius z.
	Manzaki subst.
Sematian st.	*Erodona* z.
	Kami-iwahasi subst.
	Azu subst.
	Kami-izumi subst.
Kanôzanian st.	Hitomi subst.
	Itiziku subst.
Sasage b.	

H. Ueda (1930)

Kantô loam	
Narita g.	Narita f.
	Kiyokawa f.
	Sanuki f.
Umegase g.	Kakinokidai f.
	Kokumoto f.
	Umegase f.
Kurotaki g.	Kiwada f.
	Kurotaki f.

H. Ueda (1933)

Kantô loam	
Narita g.	Narita f.
	Kiyokawa f.
	Sanuki f.
Umegase g.	Kakinokidai f.
	Kokumoto f.
	Umegase f.
Kurotaki g.	Ôtadai f.
	Kiwada f.
	Kurotaki f.

T. Mituti (1933a & b)

Tatikawa st.	
Narita st.	Kiorosi subst.
	Manzaki subst.
Yatomi st.	Kami-iwahasi subst.
	Ôyaru subst.
	Ôdorii subst.
Semata st.	Kami-izumi subst.
	Azu subst.
	Mariyatu subst.
Kongôti f.	
Kasamori f.	
Mandano sand	
Tyônan f.	
Itiziku sand	Kakinokidai f.
Kokumoto f.	
Umegase f.	
Ôtadai f.	
Kiwada f.	

N. Ikebe (1937) and K. Suzuki (1937)

Musasino system	Upper part	Narita ser.	Tatikawa st.	
			Manzaki st.	Kiorosi subst.
				Kami-iwahasi subst.
			Semata st.	Kami-izumi subst.
				Azu subst.
			Turumai st.	Yabu subst.
				Zizôdô subst.
				Mandano subst.
	Lower part	Satomi ser.	Kakinokidai st.	Kosikiya subst.
				Kawayatu subst.
			Akimoto st.	Kokumoto subst.
				Asôbara subst.
		Nokogiri-yama ser.	Nisihata st.	Sakahata subst.
				Kiwada subst.
			Seki st.	

K. Suzuki & N. Ikebe (1938, MS) -F. Takai (1938)

Musasino system	Upper part	Narita ser.	Tatikawa st.	
			Manzaki st.	
			Semata st.	Azu subst.
				Zizôdô subst.
			Sanuki st.	Kasamori subst.
				Mandano subst.
			Kanôzan st.	Kosikiya subst.
				Kawayatu subst.
	Lower part	Umegase ser.	Kokumoto st.	
			Higasi-higasa st.	
			Yûgi st.	
		Kurotaki ser.	Takagohata st.	
			Kiwada st.	
			Katuura st.	

Table II. Different opinions concerning the stratigraphic succession of the district of Ônuki-mati, Sanuki-mati and Minato-mati, Kimitu-gun, Tiba Prefecture, South Kantô, Japan.

Abbreviations:—ser.=series; st.=stage; subst.=substage; b.=beds; shell-b.=shell-beds; g.=group; sg.=subgroup; f.=formation

T. Ueji (1927)	J. Makiyama (1929)	J. Makiyama (1930a)	
Kimitu f.	Kimitu b.	Kanôzan b.	Hitomi st.
			Itiziku st.
		Sasage b.	
Yosino f.		Miura b.	
?			
Nokogiriyama f.			

J. Makiyama (1930b)			J. Makiyama (1931)	
Kimitu ser.	Kanôzan b.	Hitomi st.	Kanôzanian st.	Hitomi subst.
		Itiziku st.		Itiziku subst.
	Sasage b.		Sasage b.	
Miura ser.	Yosino b.			
	Nokogiriyama b.			

H. Ueda (1930)	
Narita g.	Kiyokawa f.
	Sanuki f.
Umegase g.	Kakinokidai f.
	Kokumoto f.
	Umegase f.
Kurotaki g.	Kiwada f.
	Kurotaki f.

H. Ueda (1933): Type section		South-western coast of Province of Kazusa	Koitogawa basin
Narita g.	Kiyokawa f.	Hitomi fossil-shell-bearing b.	Yaehara fossil-b.
	Sanuki f.	Sanuki f.	Ôyatu fossil-shell-bearing sand and gravel
Umegase g.	Kakinokidai f.	Greatly cross-bedded sandstone	Greatly cross-bedded sandstone Kanôzan shell-b.
	Kokumoto f.	Thick beds of shale	Shale
	Umegase f.	Yokoyama sand and gravel	Higasi-higasa shell-b.
Kurotaki g.	Ôtadai f.	Tômiya fossil-shell-bearing b.	Nisi-higasa shell-b.
	Kiwada f.		Anahara shell-b.
	Kurotaki f.	Ôtukayama fossil-bearing tuff	Yarazuka shell-b.

Y. Otuka (1932)		Y. Otuka & K. Motizuki (1932)	
Kanôzan st.		Kanôzan f.	Gotanda mud
			Ôyatu sand
			Sawanoyatu mud
			Itiziku sand
Akimoto g.	Awagura f.	Akimoto g.	Awagura mud
	Hirata f.		Hirata alternation of sand and mud
	Higasi-higasa f.		Higasi-higasa mud, sand and gravel
Seki g.	Takamizo f.	Takago-hata g.	Takamizo mud
	Takagohata f.		Takagohata alternation of sand and mud
	Seki f.	Seki g.	Seki tuffaceous sand
			Miyohara sandstone and conglomerate

Y. Otuka (1949)	
Sasage group	
Iwasaka fine sand	
Kiyosumi g.	Tômiya tuffaceous sandstone
	Takeoka tuff-breccia

N. Ikebe (1948, MS)—A. Huziwara & S. Ogose (1952)	
Narita g.	Semata f.
	Zizôdô f.
Turumai g.	Sanuki f.
	Kanôzan f.
Akimoto g.	
Nisihata g.	Kiwada f.
	Katuura f.

N. Ikebe (1954)		
	Narita g.	Semata f.
Miura g.	Turumai sg.	Sanuki f.
		Kanôzan f.
	Akimoto sg.	Kokumoto f.
		Umegase f.
	Nisihata sg.	Kiwada f.
		Katuura f.

K. Koike (1949)			
Kanôzan st.	Miura g.	Akimoto sg.	Itiziku sand
Umegase st.			Awagura mud
			Hirata alternation of sand and m
			Higasi-higasa sand and gravel
			Iwasaka fine sand / Sakahata alternatio
			Takamizo mudstone
Nisihata st.		Seki sg.	Komaba tuff-breccia
			Tômiya tuffaceous sandstone
			Takeoka tuff-breccia

Zonal Arrangement of Residual Clays

By

Takao SAKAMOTO

Abstract

1) Deep weathering:– Montmorillonite is formed in deep horizons in contact with a small quantity of water in pore space, very highly concentrated with cations. Beidellite and illite are formed by cation-exchange, accompanied by Al substitution in Si-O tetrahedra and Al addition in Al-O-OH octahedra, thus resulting in an Al enrichment within the lattice. The medium is alkaline ground water in phreatic motion. The clay minerals of deep weathering (tot-sheet) are supposed to be formed and transformed by ionic substitution within the lattice without complete breakdown of the structure of parent minerals.
2) Surface weathering:– Continued saturation, desaturation and leaching with freely circulating water cause unbalance of charge and "degradation" of the lattice. Kaolinite, Halloysite, hydrated halloysite (all to-sheet) and gibbsite (o-sheet) seem to be the products of isoelectric precipitation of electrically charged sols.
3) Controlling factor:– It seems to be the nature and concentration of cations that control the crystal lattice of residual clay minerals. Parent rock compositions are important in deep weathering and hydrological conditions in surface weathering.

CONTENTS

Introduction

ROSS and HENDRICKS (1945) state that "alkaline feldspars and micas tend to alter to kaolin minerals, whereas ferromagnesian minerals, calcic feldspars, and volcanic glass commonly alter to members of the montmorillonite group[1]". JENNY (1950) states that "the specific conditions which control the formation of the various clay minerals in soils are not well known. ......The roles played by climate, time of weathering, vegetation, and drainage conditions have not been elucidated[2]".

Studies on the zonal arrangement of clay minerals in nature seem to point to the fact that clay minerals adapt themselves to even subtle variation of character of medium with which they are in contact. The character of medium is partly inherent from parent rocks and partly acquired from the environment.

The writer tries, in this paper, to trace the vertical variation in character of medium along the profile of residual clays, and the corresponding variation in clay minerals, with the view to get at factors controlling clay mineral formation. The problem whether the parent minerals undergo complete breakdown into individual polyhedra and gels, or only ionic substitution and transformation within the lattice are responsible for the formation of new clay minerals, is also discussed.

1) Clay minerals in zonal arrangement in beds of sediments and on profiles of residual clays

Clay minerals in beds of sediments sometimes form concentric shells of an oolite, from fractions of one mm to a few mm thick, or thin parallel bands of nearly the same order of thickness piled one upon the other to form beds, often a few scores of meters thick. Some of the examples are:

Pre-Cambrian banded iron ores[1]
Pre-Cambrian chamosite iron ores[2]
Permian aluminous shale[3]

Jurassic oolitic iron ores[4]

On a vertical profile of residual clays on weathered outcrop of rocks, different types of clay minerals are developed at different horizons. Such zonal arrangement on soil profiles has been studied by soil chemists. Recent observations on clay and bauxite deposits have revealed clear cut picture of such zonal arrangement in many clay and lateritic bauxite deposits.

Soil profile
Fullers' earth deposits[5]
Kaolinite 〃 [6][7]
Lateritic bauxite 〃 [8][9][10]

In sedimentary beds, zoning is due to frequently repeated precipitation of clay minerals with different compositions. While, the zoning on a profile of residual clay shows a single sequence, and not a repetition, of change in clay mineral composition from bottom to top. The zoning is observed in bands of thickness from less than one m on ordinary soil-, up to over 20 m on laterite profile.

Iron rich clay minerals in beds of sediments are thought to be authigenic. So, it is inferred that the zoning in such sediments may be due to frequently repeated change in nature of medium of deposition. While the zoning shown by a residual clay profile is supposed to be due to the vertical variation in nature of medium of deposition, being more or less constant at each horizon of the profile. Such zoning is due to regular arrangement of clay minerals newly formed in situ as the product of secular change at each one of the successive horizons.

2) Profile

Acid earth deposits in northeastern Japan are usually found on the basement of liparite. The acid earth or H-montmorillonite is found near the surface, and alkaline bentonite down below[1]. Although weak hydrothermal action is suspected to have caused initial alteration, the horizontal arrangement of deposits along certain level of hillsides and the microtopography strongly suggest the action of surface and underground water to be chief agency of producing and maintaining acid earth on top of bentonite.

Hydrated halloysite is often found to be associated with montmorillonite in decomposing volcanic ash, with increasing amount of

montmorillonite in deeper horizons(2). Laterite is formed upon weathered gneiss and basalt in India(3), and it is on hornfels of Triassic age in Bintan Island(4). These are so-called "vollstaendiges Laterit", containing bentonitic clay, kaolinitic clay and bauxite in an ascending order.

These, together with other examples of kaolinite deposits, etc., were summarized in the former section of "isosials(5)". In the present paper, new "isosials" are shown in somewhat modified form in the middle of Fig. 1. Moreover, they are extended to the left to include kaolinite and gibbsite formation from such alkalic rocks as phonolite and nepheline syenite.

The bauxite in Bintan Island rests on top of a laterite which is 54 m thick(6). Other examples of lateritic bauxite, however, do not always cap thick laterite profiles. Among large bauxite deposits in the world, there are cases in which part of the bauxite consisting in gibbsite, rests directly on parent rocks. For example, in such ores as the "granitic ores" on nepheline syenite in Arkansas, U.S.A.(7), bauxite on phonolite in Brazil(8) and British Guiana(9), etc., the boundary between the bauxite and fresh parent rock is very abrupt, and the change is said to take place in a width of about 1 mm.

So, the section of "isosials" by the writer is extended to include

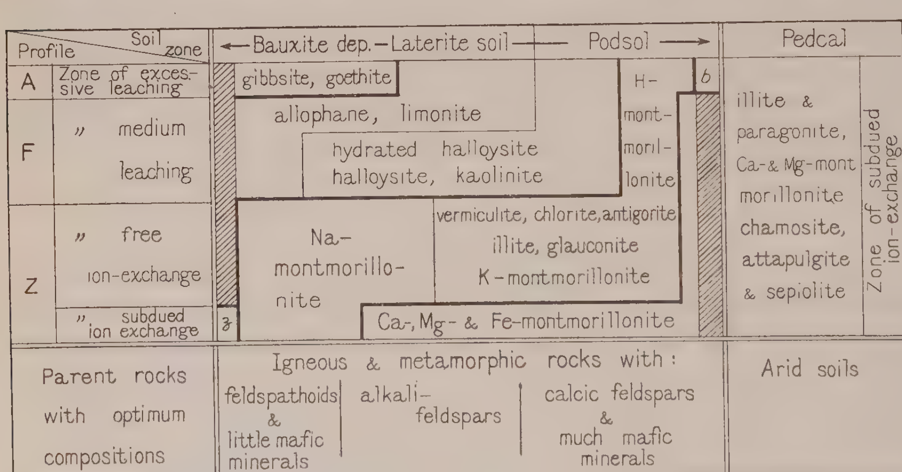


Fig. 1. Zonal arrangement and variation of environmental condition on the profile of residual clays.

Remarks: z—zeolite (analcime, etc.), b—bauerite ($SiO_2 \cdot aq$)
thick lines—"isosials".

these cases, and is presented in the new section in Fig. 1. The nepheline syenite in Arkansas, exposed on top of low hills, was extensively weathered in the Wilcox age of the Eocene epoch. The syenite was replaced by kaolinite in some parts and by gibbsite in others. The original rock texture is still maintained with these newly formed minerals pseudomorphous after feldspars. The surface of kaolinite mass, including that of redeposited kaolinite, is further changed into gibbsite in a body of blanket form.

In the extreme left of Fig. 1 is included a column of zeolite. The upper surface of the whole "isosials" represents the humid soil zones from laterite to podsol as before. On the right hand side, a column for pedcals is added, in which no vertical zoning is indicated. Below the "isosials" in Fig. 1 are shown those parent rocks which are most commonly found to be associated with respective clay minerals on the bottom of the "isosials".

3) Environmental conditions at different horizons

According to Harrassowitz, the normal laterite profile is divided into following horizons(1):

A.....Anreicherungsznoe
F.....Fleckenzone
Z.....Zersatzzone

The zone A is characterized with the surface capping of porous bauxite or iron ores. The zone Z consists in "lithomarge" in India and "tough clay" in Bintan Island, often mixed with fragments of half decomposed basement rocks near the bottom. The zone F is intermediate clay stained red and purple with irregular mottled appearance. The total thickness from A to Z of the profile amount to about 20 to over 50 m.

According to Fox and others who studied the profile along the cliff edge of the Deccan Plateau, spring issues near the top of A in rainy seasons, and at the bottom in dry seasons(2). So, A occupies the space between the highest and the lowest ground water table, and A is the zone of the freest water circulation. F is situated immediately below the permanent water table and at the zone of phreatic water where the ground water flows more or less freely according to the gradient. Z is the zone of compact clay and the water circulation is restricted. In the upper part of Z, toward the boundary between Z

and F, however, the ground water becomes gradually phreatic and, though very slow, in lateral motion.

Each zone has quite distinctive features with regard to the manner of ground water circulation.

Zone	State of ground water	Chemical reaction
A:	Dilute, oxygenated solution in rapid circulation	Excessive leaching
F:	Dilute, oxygenated solution in circulation	Moderate leaching
Z { upper part:	Concentrated solution in circulation	Free ion-exchange
Z { lower part:	Concentrated solution restricted in circulation	Subdued ion-exchange

4) Decomposition of rock-forming minerals

a. Order of susceptibility

Common products of weathering from rock-forming minerals are:

Olivine—Iddingsite, serpentine, talc, Mg- & Fe hydroxides and carbonates

Pyroxene—Uralite, chlorite, serpentime, talc, titanite, rutile, limonite and carbonates

Amphibole—Chlorite, vermiculite, limonite

Biotite—Limonite, goethite, vermiculite, chlorite, montmorillonite, kaolinite

Feldspars—Kaolinite, sericite, illite, chlorite (addition of Mg), zoisite-epidote like minerals, zeolite, calcite

Among the rock-forming minerals, feldspathoids, olivine, Ca- and Na feldspars are more easily decomposed than pyroxenes, amphiboles and biotite, while potash feldspar, muscovite and quartz are most resistant. GOLDICH once showed the sequence in a diagram very much resembling that of the reaction series by N. L. BOWEN[1]. The higher the position on the series, the easier get decomposed the minerals. The minerals on the series show the following sequence in their crystal lattice:

Mafic minerals:		Salic minerals:	
Olivine	SiO_4	Feldspars	$SiO_{4/2}$ – three dimensional
Pyroxene	Si_2O_6 – chain		

Amphibole Si_4O_{11} – band
Mica Si_2O_5 – sheet

In mafic series, it appears to be that the lower the polymerization of Si-O tetrahedra, the more unstable are the minerals. In salic series in which all minerals have three dimensional lattice of $SiO_{4/2}$, the higher the substitution of Al for Si, the less stable are the minerals.

b. Hydration and hydrolysis

During the weathering process, water-free silicates become hydrolyzed to hydroxy-silicates; those silicates with less OH become replaced by those with more OH. Simple addition of OH to augite chain structure easily changes it into hornblende band structure. Chain and band structures are polymerized and silicified to sheet structure:

$$\underset{\text{orthosilicate ion}}{4(SiO_4)^{4-}} + 6H_2O \rightarrow \underset{\text{sheet anion}}{(SiO_{4/2})_3SiO_4^{4-}} + \underset{\text{hydroxyl}}{12(OH)^-}$$

Three dimensional silicate with $(Si, Al)O_{4/2}$ is decomposed to compounds of sheet type:

$$\underset{\text{three dimensional alumosilicic acid}}{2(SiO_{4/2})_2(AlO_{4/2})_2^{2-}} + 6H_2O \rightarrow \underset{\text{sheet anion}}{(SiO_{4/2})_3SiO_4^{4-}} + \underset{\text{sheet like precipitate of hydroxide}}{4Al(OH)_{6/2}}$$

Na^+, K^+ and Ca^{2+}, or elements of the group I (see Chapter 5 b) and later Mg^{2+} and Fe^{2+}, or elements at about the boundary between groups I and II, go into solution as cations. While Al^{3+} and Fe^{3+}, being elements of group II, first form colloidal sol but soon get flocculated to form more coarsely dispersed hydroxide and get precipitated.

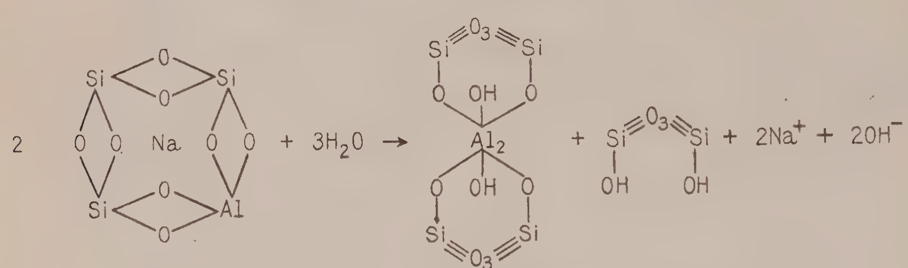
Feldspars are the most important of all the rock-forming minerals which are altered into clay minerals upon decomposition. The structure of feldspars and feldspathoids is shown by three dimensional networks of $SiO_{4/2}$, being modified by the substitution of Al^{3+} for part of Si^{4+} and addition of such cations as K^+, Na^+ and Ca^{2+} for balancing the charge deficiency caused by the substitution.

$$4SiO_{4/2} \quad \text{to} \quad \underset{\text{orthoclase}}{[(SiO_{4/2})_3AlO_{4/2}]K} \quad \text{or} \quad \underset{\text{anorthite}}{[(SiO_{4/2})_2(AlO_{4/2})_2]Ca},$$
$$\underset{\text{albite}}{[(SiO_{4/2})_3AlO_{4/2}]Na} \quad \text{or} \quad \underset{\text{nepheline}}{2[SiO_{4/2}\cdot AlO_{4/2}]Na}$$

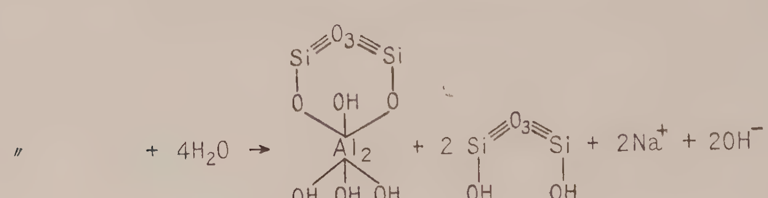
Such feldspars react with water and become essentially unstable under condition of dispersion.

According to FREDERICKSON (1951), the hydration and hydrolysis are explained by the process of ion-exchange[2]. When exposed to water or water vapor, a film of polarized water is adsorbed to the crystal surface of, for instance, albite. Then by an ion-exchange, H^+ goes into the crystal lattice and Na^+ goes into the water layer. This exchange causes the strain and expansion of crystal lattice so that the exchange may keep on going into the crystal. Although FREDERICKSON's idea seems to have not gained unanimous support in minor detail, the writer believes that it is probable that hydration and hydrolysis rather than solution by acid are responsible for the incipient decomposition of feldspars, because the reaction proceeds under alkaline condition far below the ground water table.

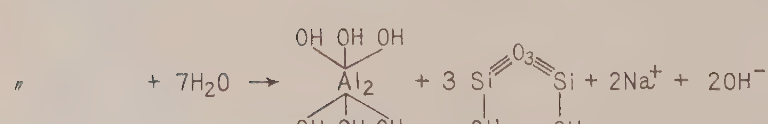
Such expanded and unstable lattice would absorb more water until finally Al and Si become hydroxides. The hydrolysis of an albite crystal may be shown schematically as follows:



albite montmorillonite silica gel



kaolinite silica gel



gibbsite silica gel

The environment of weathering is one of oxidation. Under the temperature of weathering, OH or H_2O addition is the first result of transition of Fe^{2+} to Fe^{3+}. The bridge oxygen is broken after the scheme of $O + H_2O \rightarrow 2(OH)$, and OH group enters in place of the O. Al^{3+} and Mg^{2+} as well as Fe^{3+} eventually go into hydroxide and oxy-hydroxide.

The sequence in which the lattice is hydrolyzed, or whether new clay minerals with definite crystal lattice is directly formed or the original crystal lattice is destroyed into separate Al and Si gels to recombine later into clay minerals is a question(3). CORRENS and ENGELHARDT(4) are of the opinion that weathering proceeds by the breakdown of the parent mineral to ionic solutions or gels and secondary minerals are the reaction products. GRIM states that the transition from a primary mineral or minerals to a clay mineral weathering product may not always take place in exactly the same way, but it would depend on the parent mineral, the resulting clay mineral and the environmental conditions(5).

NIGGLI, too, states that it is possible for the transformation to take place without complete breakdown of the parent mineral structure in some cases and with breakdown in other cases(6).

Taking this point into account, a suggestion by the writer will be stated in the following chapters regarding the sequence of clay mineral formation. The sequence is inferred by the writer upon the basis of observed parallelism between the chemical and crystal structural changes of clay minerals on the one hand, and changes in environmental conditions which control the nature of medium at different horizons of the natural profile on the other.

5) General properties of clay minerals

a. Structural features

Most clay minerals are so-called layer minerals whose structural units are:

octahedral sheet, o: $(X, Y)_{2-3}(OH)_6$
tetrahedral 〃 , t: $(Z_2O_3)(OH)_2$
〃 band, $(Z_4O_7)(OH)_4$

These units are combined to form
$(X, Y)_{2-3}(Z_4O_{10})(OH)_2$ or tot three sheet layer;

when adsorbed cation (W) and interlayer water are present,

$(X, Y)_{2-3}(Z_4O_{10})(OH)_2 \cdot mH_2O(W_W)$, and
$(X, Y)_{4-6}(Z_4O_{10})(OH)_8$ or to two sheet layer, or
$(X, Y)_{12-14}(Z_{16}O_{44})(OH)_4$ double chain or band.

X: Al, Fe^{3+}, Mn^{3+}, Cr^{3+} (two of the octahedral positions occupied and the third vacant—dioctahedral series),
Y: Mg, Fe^{2+}, Mn^{2+}, Ni, Zn, Li, Ti (all three octahedral positions occupied—trioctahedral series),
Z: Si, Al,
W: K^+, Ca^{2+}, Na^+.

As long as Z is composed of only Si, the outer surface of t sheet has only O of complete double bond, and the charge is neutralized. When Al is substituted for a part of Si, then part of O has free bond directed outward. Then it follows that either such cations as K^+, Ca^{2+}, Na^+ move into the space between layer (W) or X is substituted for a part of Y, thus the free bond of O is satisfied. Sometimes o sheet completely filled with hydroxyl enters between tot layers. There are different ways of piling up one and two sheet layers one above the others. Especially in so-called active clays such as montmorillonite, interlamellar adsorption (W) due to the lattice substitution is known by its strong ion-exchange and swelling.

Clay minerals have the characteristic monotonous habit of silicates with sheet structure, showing pinacoidal cleavage and lamellar forms. The most important feature is that clay minerals show a regular but extreme complexity caused by the variation of composition due to the lattice substitution, and by the presence of mixed layer minerals.

b. *Chemical features*

Clay minerals consist chiefly in elements Si, Al, Mg and Fe, with minor amount of K, Ca and Na. According to the dominant element beside Si, they are divided into aluminian (dioctahedral)-, magnesian (trioctahedral)- and ferriferrous series.

Geochemical considerations have been given to the reason why these elements are apt to remain in clay minerals on the earth's surface. GOLDSCHMIDT[(1)] divided elements into three groups by ionic potential.

Ionic potential $\phi = Z/r$, Z-valence, r-ionic radius

	Ions:	ϕ:		Ions:	ϕ:		Ions:	ϕ:
			Group II	Si	10.26			
				Mn^{4+}	7.69	Group III	N	33.33
Group I	K	0.75		Fe^{3+}	4.48		C	~26.7
	Ca	1.89		Al	5.26		S	17.65
	Na	1.02		Mg	2.56		B	~15
				Fe^{2+}	2.41		P	14.29
				Mn^{2+}	2.19			

GOLDSCHMIDT observed that the elements of Group I are held in ionic solution, while the elements of Group II form hydroxides by hydrolysis and are precipitated in sediments (hydrolyzates). Those of Group III are combined with oxygen to form negative ions or acids.

Later WICKMAN[2] gave the above empirical threefold division a physical explanation by replacing the concept of the ionic potential with rules governing the hydrogen and hydroxyl bonds in hydroxides. He used electrostatic valence;

electrostatic valence $V=Q/C$, Q-charge of cation, C-co-ordination number of cation

$V<1/2$, ionic bond in hydroxides, very soluble (alkaline solution) ....Gr. I

$1/2\leqq V\leqq 1$, hydroxyl bond in 〃 , precipitated as hydroxides ....Gr. II

$V>1$, hydrogen bond in 〃 , very soluble (acid solution) ....Gr. III

Those elements which form hydroxyl bond are precipitated as hydroxides. The 6-co-ordinated trivalent cations such as Al in gibbsite: γ-$Al(OH)_3$, diaspore: α-AlO(OH), and Fe^{3+} in goethite: α-FeO(OH), lepidocrocite: γ-FeO(OH) are the most typical of the Group II. The 4-co-ordinated quadrivalent cations such as Si and Mn^{4+} also belong to Group II.

Cations of Group I are adsorbed to clay minerals from their ionic solution. They are only in minor amount in clay mineral constituents. But they form alkaline solutions which causes the unstability of Si-O tetrahedral structures. Moreover, their adsorption and ion-exchange cause the lattice substitution. They play much more important role in clay formation than they appear at first sight.

Clay minerals contain less hydroxyl when their composition is close to the parent mineral. The less the content of Si (t/o), the larger the amount of hydroxyl and the less the capacity of ion-ex-

change. The amount of hydroxyl and the nature and amount of adsorbed cations, therefore, are in an intimate relation to the composition of clay minerals. To what extent and how both of these are incorporated in the clay mineral is the point to be elucidated in the problem of clay genesis.

It is the writer's intention to try, in this paper, to trace the transformation of clay mineral to the interaction between the so-called "hydrolyzates" of the Group II elements and ionic solutions, especially alkaline solutions of the Group I elements.

6) Sequence of clay mineral formation in the process of weathering in the light of the observations of natural profile of residual clays

a. Restricted hydration and ion-exchange

(1) Subdued ion-exchange and formation of montmorillonite

Ca-montmorillonite and effect of Ca^{2+} ion

When feldspar, for instance, is first hydrolyzed either underneath a thick cover of weathering product, as deep as 50 m in some cases, or confined in thick volcanic ash deposited and subjected to gradual compaction in a bed on the bottom of shallow water, very little quantity of water is circulated within the mass which is being decomposed. So, the small amount of liquid water in the pore space would become very highly concentrated solution with cations liberated in situ.

It is known in different cases that high salinity tends to cause dehydration. So, there are very small amount of OH available for newly formed clay minerals. The ratio $OH/(O+H)$ in tot layer is 2/12, whereas the ratio in to layer is 4/9. It would be easier for tot than for to layer to be formed under condition of restricted supply of OH. It is supposed that the supply of water in the newly formed mass of montmorillonite adjacent to feldspar crystal is so small that, except the water molecules taken into interlayer water, the remaining amount of liquid water is enough to cause only a part of Si-O tetrahedra dispersed into sol and taken away from the system. In other words, the water supply would not be enough to disperse all the Si-O or Si(Al)-O tetrahedra into separate SiO_2 and Al_2O_3 sols. The concentration of cations is also too high—the more the decomposition proceeds, the higher it becomes—and is above the threshold value for flocculation. Thus SiO_2 transported may be in an ionic solution instead of sol.

The high concentration of cations makes easy the adsorption of these by clay minerals. The clay minerals formed under such condition of restricted water circulation and least leaching are montmorillonite, saturated with cations available in situ.

When plagioclase feldspar is decomposed much Ca^{2+} is liberated. Ca^{2+}, when adsorbed, tends to make the montmorillonite lattice very firm. The charge of cations adsorbed by montmorillonite is said to be chiefly balanced by Y exchange in octahedra(1). If Al^{3+} of Al-O-OH octahedra is replaced by Mg^{2+}, Ca^{2+} is adsorbed to make the lattice very firm. In other words, the presence of Ca^{2+} tends to maintain SiO_2/Al_2O_3 ratio high, or it is in opposition to Al enrichment. It is also shown by hydrothermal synthesis that montmorillonite is best produced in the presence of Mg^{2+} (2).

Na-montmorillonite and effect of Na^+ ion

When feldspathoids and feldspars rich in alkalies, such as in phonolite, nepheline syenite and also in granite are decomposed, the first product seems to be Na- instead of Ca-montmorillonite. This is inferred by tne observation of highly swelling bentonite beds among beds of marine sediments. The Ashiya formation in northern Kyushu coal-fields contains a few beds of swelling bentonite. When met with in underground tunnels they cause much damage even to thick concrete construction. This swelling is thought to be due to Na-montmorillonite derived from acidic ashes.

Chemical analysis of a swelling bentonite from Ashiya beds:

SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	Mn	MgO
57.87	0.29	15.19	2.43	1.00	0.011	1.63
CaO	Na_2O	K_2O	$H_2O_{(-)}$	$H_2O_{(+)}$	Total	
0.87	2.40	0.32	13.85	3.92	99.78	

Analyst: Hiroshi HARAMURA.

The Na-montmorillonite is thought to have been formed within the bed of ashes in contact with small amount of water in pore space concentrated with Na^+ and left to stand intact within the bed. It had never been deflocculated in sea water to get Na^+ exchanged with other cations such as Ca^{2+} in sea water. Otherwise non-swelling Ca-montmorillonite would have been formed, because it is a more natural

product of flocculation in sea water.

Any Na-montmorillonite produced upon the surface of alkali rich parent rocks is not enclosed and protected by overlapping beds of sediments, but is exposed to circulating water. Deflocculation and transportation make the path easier for further circulation of water, and hence deflocculation and transportation of more Na-montmorillonite.

In the case of Ca-montmorillonite, as is usually the case, it is practically not dispersible and forms a thick mass covering the parent rock in situ. Granites weather into arkosic sands very easily, while basic rocks, dolerite, gabbro and some diorites weather into large and small balls embedded in mass of clay. This may depend upon the difference in the capacity of deflocculation of the first product of weathering. Ca-montmorillonite is transformed into clay minerals easily dispersible and more susceptible to environmental change first in contact with slowly circulating ground water and subject to ion-exchange in an upper horizon.

When the parent rock is alkalic, especially when rich in Na^+, the first product is Na-montmorillonite. It is highly dispersible(3) and can form a mass of appreciable quantity only in such a confined space, sealed against circulating water, as in the bed of volcanic ashes deposited on water bottom and sealed by later sediments piled above it. The fact that the liquid medium within the bed of ashes is often very strongly alkaline is shown by the occurrence of zeolite in such beds. In one of the oil-shale beds with tuffaceous materials of the Green River formation in Utah and Colorado, U.S.A., analcime crystals are found to form 16% of the total mass of the bed(4).

Thus the presence of the Na^+ tends to form, in most cases, highly dispersible and unstable Na-montmorillonite and, though in rare cases, even zeolite. Na-montmorillonite, in contrast to Ca-montmorillonite, adsorbs so much interlayer water that the layers finally become randomly dispersed in water.

Zeolite has three dimensional tetrahedra like the parent feldspar crystals, and is quite an exceptional case among clay minerals. This may be related to the strong power of Na^+ to deflocculate Si-O tetrahedra.

(2) Active ion-exchange and transformation of montmorillonite

A little higher above along the profile, freer circulation of ground water causes the change in nature and concentration of cations in

solution at a given place. Then an active exchange of adsorbed cations takes place. The exchange rate varies with the nature and concentration of cations, as is well shown by the fact that an artificial water softener can be repeatedly used for many times after treatment with saturated solution of NaCl.

Frequent exchange of cations causes the unbalance within the lattice. Although Y exchange (substitution of Mg^{2+}, Fe^{2+} etc. for Al^{3+} and Fe^{3+} in octahedra) is normal for montmorillonite, the high concentration of alkali ions seems to cause Z exchange (substitution of Al^{3+} for Si^{4+} in Si-O tetrahedra up to 1/6 in illite)[1], too. This seems to be the case especially with K^{+} ion and to some extent with Na^{+}. Repeated exchange of adsorbed cations, including H^{+}, make the lattice of clay minerals become unbalanced. Slight exchange of Al^{3+} for Si^{4+} in tetrahedra may be internally balanced by addition of Al^{3+} to octahedra or replacing OH for O in octahedra[2]. This may be one way of Al enrichment in the lattice.

In such unbalanced and so-called degraded form, some of the Al^{3+} is in exchangeable position and it is shown by experiments that large amount of cation can be adsorbed. As the result of K-fixation, clay mineral lattice become very similar to mica or illite. The beidellite is supposed to be a layer by layer mixture of illite and montmorillonite. So the change montmorillonite→beidellite→illite is an important course of Al enrichment. If Na^{+} takes part, paragonite is formed in place of illite. This reaction seems to take place in the alteration in spilite formation.

Thus Na-montmorillonite, formed either by an ion-exchange from Ca-montmorillonite or by direct decomposition of parent rock-forming minerals, is dispersed in solution of varying concentration of ions in the upper part of the Zersatzzone. It is subjected to the change above stated. It follows the course of Al enrichment during the alteration due to ion exchange.

High concentration of Na^{+} tends to deflocculate SiO_2 gel into solution and cause it to be taken away from the system. The role of alkali ion is, therefore, two fold : the formation of clay mineral lattice with higher Al and the removal of freed SiO_2. The role of K^{+}, being easily fixed in the lattice to cause widespread formation of illite, seems to be more important in the first reaction, whereas the role of Na^{+} in the second.

The reactions stated in the above (1) and)2) can be summarized

as follows:

If the parent mineral is rich in Ca^{2+}, Ca-montmorillonite is formed and remain stable but is later gradually changed into less stable form as the result of ion-exchange. If the parent mineral is rich in alkalies, alkali-montmorillonite or unstable form is directly formed. All of these unstable form of clay minerals are subject to further change, resulting in increased Al substitution for Si. Removal of stabilizing Ca^{2+}→formation of dispersible Na^{+}-montmorillonite→degradation of crystal lattice→(adsorption of K^{+} to form illite) seems to be the normal course taking place in the so-called Zersatzzone, or zone of decomposition. The kind of clay minerals or their crystal lattice, seems to be controlled by the amount of available OH and nature and concentration of cations.

b. Intensive hydration and leaching with isoelectric precipitation

(1) Leaching of cations and unbalanced lattice

Such cations as Ca, Mg, Ba, Li, Na, K, and Cs within the lattice can be brought to exchangeable position by acid leaching or electrodialysis(1). By continued leaching or desaturation of montmorillonite, Al and Fe are known to move into exchangeable position or clog the path of cations to be exchanged. This may be interpreted as the results of unbalance of charge within the lattice due to rapid elimination of cations and stronger hydration (substitution of OH for O) of octahedron. This is why H-montmorillonite is said to be actually a H-Al system(2). When alkalies are leached out of an illite mineral, the resulting charge deficiency may be balanced within the lattice by adding Al to the octahedra, increasing the total Al more than the normal 2. This is another course of Al enrichment. But such Al may pass into exchangeable position when leaching continues.

Such is the degraded condition of crystal lattice of montmorillonite as well as mica like minerals.

(2) Formation of kaolinite and halloysite

H-montmorillonite may be reckoned as a low temperature equivalent of pyrophyllite and is fixed only under strongly acid condition. Under nearly neutral conditions with ample water circulation, Si-O tetrahedra are dispersed in random direction and Al-O-OH octahedra are more highly hydrated. And the resultant lattice is to instead of less hydrated

tot. to is the lattice of kaolinite, but incomplete lattice with randomness of spacing and interlayer water is halloysite and hydrated halloysite.

Si, being one of the elements of Group II, is hydrolyzed to take OH and get precipitated. But, being situated on the border of Groups II and III elements, it also combines, like those of Group III, with O and become an anion. This is why Si-O tetrahedra is less easily hydrolyzed and dispersed but very easily fixed in the presence of H^+ or under acid condition, while it is easily hydrolyzed and dispersed in neutral to alkaline media. When kaolinite or halloysite is in contact with ample water of alkaline reaction, more of t is eventually hydrolyzed and deflocculated out of to, until finally all the t is lost, leaving behind only o or gibbsite if hydration continues.

Al, like Fe^{3+}, belonging to Group II, forms a stable hydroxide in colloidal form. Being situated in the middle of Group II, they are not neutral but "amphoteric" colloids, that is, behaving like positive colloids in acid medium and negative ones in alkaline. So, they move in solutions both strongly acidic and alkaline, but they are best precipitated under neutral to weakly alkaline conditions or conditions that obtain under soubtropical or tropical regions.

The electrical charge is very well balanced in kaolinite lattice without any stabilizing cations. Isomorphous substitution within the lattice is negligible and cation exchange capacity is also very small, being probably mainly related to broken bonds at the edge of crystallites[(1)]. H-montmorillonite, too, may appear at first sight to be balanced without any stabilizing cations. But H-montmorillonite is formed under strongly acid condition, in which Al-O-OH octahedra become movable, rendering the whole H-montmorillonite lattice an unstable one. This is why H-montmorillonite cannot be so common. While in kaolinite which formed under slightly acid condition, Si-O tetrahedra are not disturbed and Al-O-OH, too, are no more disturbed. First turning into alkaline condition, Si-O tetrahedra begin to be dispersed. This is why kaolinite is the most stable lattice even without any stabilizing cations and is the most widespread type of clay mineral.

(3) Formation of gibbsite

Allophane minerals are amorphous to X-ray diffraction and are thought to be mixture of gels of Si and Al hydroxides. Each of these

hydroxides is to be reckoned a deflocculated t or o sheets, that is, sheets put into random orientation.

In the degraded form of crystal lattice due to repeated saturation and desaturation, or in the weakly combined lattice with randomness in spacing, the transformation takes place in the unstable t or o sheet, which eventually are deflocculated (forced into random orientation) and leave the layer. Such deflocculated sheet is charged. So, any lattice with partially deflocculated sheets is charged, too. The larger the amount of charge, the more the lattice or particles repel each other. Thus only those lattice which have no sheet deflocculated at all, or those newly formed by combination of lattice or particles of opposite charge, become precipitated.

In other words, lattice or particles of minimum charge is fixed under a given condition. In this sense, it seems to the writer that S. MATTSON's interpretation of soil colloidal behavior with the idea of iso-electric precipitation[1] is to be credited.

The effect of continued leaching depends upon the pH. If the solution is weakly alkaline, o sheet is precipitated and t sheet is dispersed and removed. Gibbsite, thus precipitated in a large amount, forms a bauxite deposit. Close studies of mode of occurrence of bauxite, show that it is enriched by replacing kaolinite, feldspar and even quartz crystals with gibbsite. The replacement is volume by volume and the pseudomorph after feldspar laths in igneous rocks is often met with. Such replacement seems to be possible where there is ample supply of circulating water to deflocculate and remove Si-O tetrahedra. Otherwise, montmorillonite would be formed under restricted water circulation and less amount of available OH, owing to too high salinity.

SiO_2 on the other hand gets precipitated in acidic medium, where Al_2O_3 is removed. In bauxite deposits, like those in Arkansas[2], Georgia, etc., bauxite is often replaced by kaolinite, the replacing mass being in vertical funnel shape or irregular protuberances filled with kaolinite. This means the attack on bauxite by acidic water, and precipitation of kaolinite in such a place where the medium is neutral. Bauxite blanket is sometimes capped with hard and highly siliceous clay[2], which presumably is mixture of chalcedonic quartz and kaolinite. This means that the attack by acidic water was so strong as to maintain an acidic condition in a zone on top of the bauxite.

As is shown by these examples both Si-O tetrahedra and Al-O-OH octahedra can be deflocculated into gel form or in random dispersion. But they are deflocculated and flocculated respectively in an opposite condition with regard to pH of the medium and tend to get separated in either acid or alkaline condition maintained for long enough period of time for continued leaching. In neutral to slightly acid condition kaolinite remain the most stable lattice, even without any stabilizing cations.

Therefore, in the zone of leaching, the controlling factor is the pH of the medium or the isoelectric point of colloid and not the stabilizing cations.

7) Clay minerals of magnesian and ferriferrous series

The discussion in the preceding chapter is concerned with clay minerals of aluminian or dioctahedral series. Salic minerals such as feldspars and feldspathoids are naturally the more important parent minerals. There are great families of important clay minerals derived from mafic minerals—magnesian and ferriferrous series. These are especially important clay minerals in decomposed glassy tuff of Tertiary age in Japan[1]. From the structural point of view, clay minerals of different series may be grouped as follows:

Lattice	Aluminian series (dioctahedral)	Magnesian series (trioctahedral)	Ferriferrous series (dioctahedral)
tot sheet	montmorillonite	saponite	nontronite
tot 〃	illite		celadonite, glauconite
chain		sepiolite, attapulgite	
mixed tot-o sheet		chlorite, vermiculite	
to 〃	kaolinite, halloysite	amesite, antigorite	cronstedtite, chamosite
	hydrated halloysite		
gel	allophane		
o sheet	gibbsite	brucite	goethite

The stability of clay minerals of aluminian series is controlled by the availability of OH and nature and concentration of cations. That of magnesian and ferriferrous series is controlled not only by these factors but also by the oxidation-reduction potential. The reduction into Fe^{2+} of Fe^{3+} in octahedron of glauconite would increase the charge deficiency and hence the adsorption capacity for K^{+}. In chamosite the oxidation of Fe^{2+} to Fe^{3+} causes its separation into goethite and kaolinite(2). Chlorite is composed of composite layers of biotite and brucite like octahedral sheets. From the chemical composition, chlorites have been considered an isomorphous series with end members of antigorite and amesite. But X-ray analysis has shown that antigorite and amesite, having kaolinite structure, should better be dissociated from chlorite family(3). The so-called iron chlorite, like chamosite(4) is said to have kaolinite like structure.

These minerals are well developed in medium with plenty of cations and under reducing condition. Under either leaching or oxidizing condition, they are easily decomposed with the separation of kaolinite structure on the one hand and with the change eventually leading to highly hydrolyzed and oxidized forms such as brucite and goethite or limonite.

They are found in association with aluminian series in restricted ground water circulation zone in deep weathering. But in the zone of leaching they get destroyed very easily. Upon the land surface these are stable only in arid soils or in sediments of alkaline lakes in arid region or on the bottom of confined sea with restricted circulation of sea water. They are in association with aluminian series which are stable under these environments. The column of pedcal or arid soil on the extreme right hand side of Fig. 1 shows the assemblage of those minerals.

Some of the finest example of zonal arrangement is shown by these minerals in a bed of sediments. The zonal arrangement of these is better displayed with the diagenesis of sediments rather than with the residual clay formation.

8) **Summary and conclusion**

(1) The highest concentration of cations and the least amount of available OH tend to form three sheet (tot) clay minerals such as

montmorillonite. The water in the pore space within the decomposing mass in contact with the bed rock is not only small in quantity but so highly concentrated with cations that OH is available only to a limited extent for newly formed minerals. Cations are no sooner liberated from parent minerals than adsorbed by montmorillonite in situ.

(2) Ca^{2+}, when adsorbed, fixes the montmorillonite, while Na^{+} disperses it.

(3) Ion-exchange and more hydration cause the unbalance in and degradation of lattice of montmorillonite.

(4) Alkali ions, especially K^{+}, cause Al^{3+} substitution for Si^{4+} in Si-O tetrahedra.

(5) Al^{3+} addition to Al-O-OH octahedra and Al^{3+} substitution in Si-O tetrahedra cause Al enrichment in the lattice of successive series of newly formed minerals.

(6) Alkali ion, especially Na^{+}, causes dispersion and removal of Si-O tetrahedra, that is, desilication.

(7) Increased circulation of ground water leaches and removes all the cations, even Ca^{2+}, Mg^{2+}, and K^{+} from within the lattice.

(8) Si-O tetrahedra and Al-O-OH octahedra, when highly hydrated, get dispersed. These have electrical charge of opposite sign and get mutually precipitated as isoelectric precipitates.

(9) Kaolinite is stable without any stabilizing cations, because the composition corresponds with that of the isoelectric precipitate in neutral to weakly acid medium.

(10) Further hydration tends to disperse all the tetrahedra and octahedra in random orientation. They are dispersed and flocculated in media of opposite pH, eventually resulting in the separation of Al_2O_3 gel from SiO_2 gel, or gibbsite from chalcedonic quartz.

In view of the sequence of these changes, it seems to be the amount and nature of cations and also amount of available OH that control the crystal lattice in the early stages of clay mineral formation. Because these changes take place at or near the surface, it is neither temperature nor pressure that controls the formation of layers of three-, two- or one sheet, and other structures. The internal structure seems to be controlled by the medium outside.

The formation of montmorillonite from parent mineral and its transformation to illite and some kaolinite may be accomplished

directly by a rearrangement in the internal structure. Kaolinite in most part, halloysite and gibbsite are isoelectric precipitates.

Highly dispersible and hence unstable nature of Na-montmorillonite and free ground water circulation seem to account for the apparently direct formation of kaolinite, and even of gibbsite upon some granites and nepheline syenite, etc. in tropics.

Because the nature and amount of cations released depend upon the nature of parent rocks, the role of these rocks is very important especially in the bottom horizons.

Because the amount and pH of water depend largely upon the meteorological conditions the role of these is extremely important in the top horizons.

From the structural point of view, clay minerals of aluminian series are arranged as follows:

(a) montmorillonite
beidellite
illite

kaolinite

(b) halloysite
hydrated halloysite
gibbsite

The composition of parent minerals plays a dominant role for the formation of the group (a), and intense leaching and surface hydrological environments play a dominant role for the formation of group (b).

This is comparable with hydrothermal alteration in which the products are varied and complex by weak hydrothermal action, depending upon the nature of parent rocks, whereas the end products tend to become similar by intense action, almost regardless of composition of parent rocks(1).

It seems to be possible for the group (a) minerals to be formed by rearrangement of internal lattice, while the group (b) minerals seem to be formed by the complete breakdown of lattice of parent minerals and later reaction between sols, or perhaps to some extent, ionic solutions. It seems to be important to divide clay minerals into these two groups in discussing the manner of their transformation.

The motivating power to cause ion-substitution and rearrangement of lattice is the cations, their nature and concentration.

Majority of magnesian and ferriferrous series fall into group (a). They persist on the surface only in arid soils.

Clay minerals at different horizons of mature profile are in equilibrium with the surrounding medium. Therefore, the whole system is in a sort of stationary state. The zonal arrangement is maintained as long as the environmental conditions remain undisturbed.

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An Exact Treatment of the Barometer Method

By

Arata SUGIMURA

1. Introduction

In geologic field mapping one can measure the altitude of a locality by means of the barometer (e.g. LAHEE[1]). From the viewpoint of inductive statistics, the writer has treated this method somewhat rigorously. It is the purpose of this paper to give only the general course of the treatment, because two examples of application of it have been already published[2)3)].

2. Some remarks on measurement in general

How should one take the course of measurement in general ? Here the writer intends to give his answers to this question.

It is the basic principle to measure any quantity with a precision that is necessary and sufficient for the purpose of the investigation. *Therefore* measurement is to yield not only one value, but also the error for it. These are the well-known concepts among scientists.

Errors are classified into systematic and accidental ones, usually on the basis of their sources. This classification, however, is not substantial for observer, because in most cases the source of error is essentially unknown.

In the writer's opinion, whether a value is real or not, is usually clarified by coincidence of two or more values obtained through different courses of observation (independent check). If these values show a significant difference in the sence of inductive statistics, the observation is defined to have a systematic error, and if they show only an insignificant one, the difference is defined to be an accidental error.

When the observed values are accompanied by systematic errors, one should improve the method of the measurement or correct them with some factors to be found. But when they are accompanied merely by accidental errors, one can regard them as competent for

the further test to decide whether they are sufficient in precision for the purpose of the investigation or not.

3. The method of using two barometers

The writer has adopted the method of using simultaneously two aneroid barometers for the measurement of altitudes. One of the barometers is set at the base camp and is arranged to register automatically the change in atmospheric pressure there. The observer carries the other portable one and reads the atmospheric pressure at the points where he wants to measure altitudes.

From these values one can obtain the differences z's between the altitude of the base camp and those of the points in question through the following well-known LAPLACE's formula.

$$z=18400(1+0.00366t)(\log p_0-\log p)$$

where t: atmospheric temperature in °C

p_0: atmospheric pressure of the base camp in mm Hg

p: atmospheric pressure of the points in question, in mm Hg.

Corrections: Among various corrections of the values p_0 and p, the most important ones are the following.

1. Graduation correction and temperature correction

The writer adopts the mean value of the results of official approval before and after the survey.

2. Correction for the mechanical fluctuation of the barometers themselves

The fluctuation may be revealed from the interpolated values of readings of the two barometers in every morning and evening during the field surveying.

Rejection of samples: Samples under unfavourable conditions are rejected. Such cases are brought about by some mechanical causes (such as a shock) and uneven weather.

One should consider, in addition, some corrections or rejections other than these, when the observed values have a systematic error.

4. Some devices for estimation of the standard error

The writer's devices to estimate the standard error of z_i are composed of the following operations.

1. Besides the observed values $z_1, z_2 \cdots\cdots$ in question, the observer measures by the barometric surveying $z_{01}, z_{02} \cdots\cdots$ of the points where the real altitudes have been already known.

2. There should be no systematic difference in conditions between the measurement of $z_1, z_2 \cdots\cdots$ and that of $z_{01}, z_{02} \cdots\cdots$. Here the writer cites an example.

1) Condition concerning the time and date

One can expect a similarity of the conditions of weather or machine only in almost simultaneous mesurements of z_i and z_{0i}. Table 1 shows an example.

Table 1. Numbers of Times of Measurement

Date	October								November	
	16	17	18	19	20	22	25	26	22	24
z_{01}					1				1	
z_{02}				1		1				
z_{03}		1				1				
z_{04}		1	1							
z_{05}								1		1
$z_1, z_2 \cdots\cdots\cdots$	5	30			4	55	4	5	4	12

2) Condition concerning the position

(1) Range of area

The range of areal distribution of the points of $z_1, z_2 \cdots\cdots$

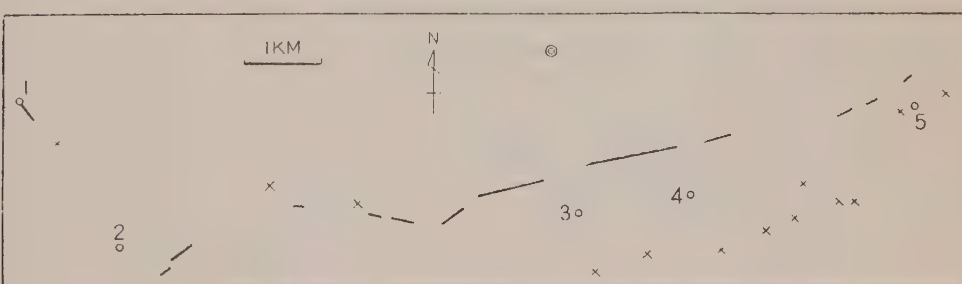

Fig. 1. Range of areal distribution of the measured points

◎: Base camp

○: Points where the real altitudes have been already known (Figures indicate the suffix of each z_0.)

×: Points in question

—: Series of the points in question

should be essentially the same with that of $z_{01}, z_{02}\cdots\cdots$. An example of this relation is shown in figure 1.

(2) Range of altitude

As is shown in figure 2, the same attention as (1) is paid about the range of altitude.

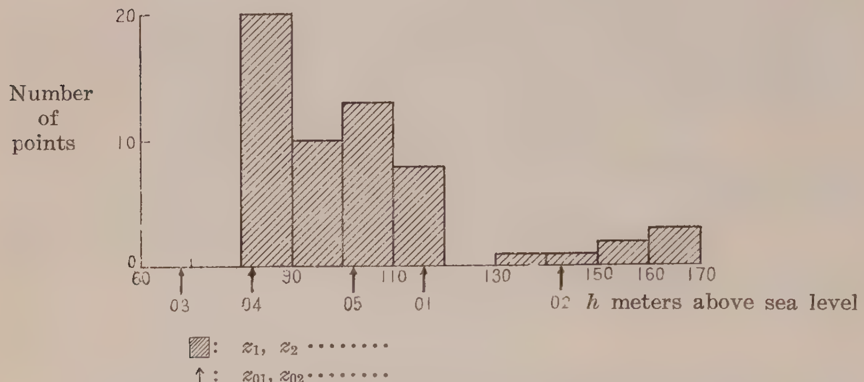

Fig. 2. Range of altitude distribution of the measured points $h=h_0+z$, where h_0 is the altitude of the base camp.

These conditions 1)·2) have to be designed before the survey.

3. Differences $\Delta z_{01}, \Delta z_{02}\cdots\cdots$ between $z_{01}, z_{02}\cdots\cdots$ and the real altitudes corresponding to them are obtained. Table 2 shows an example.

Table 2

(unit: meter)

Δz_{01}:	-1.3,	-2.8
Δz_{02}:	$+1.5$,	$+1.0$
Δz_{03}:	-0.1,	-1.1
Δz_{04}:	-1.0,	$+1.0$
Δz_{05}:	$+1.0$,	-1.5

4. One should test whether the mean value of Δz's may be zero or not, assuming their population to be normal. The following numerical values are for the example cited above.

$$\text{Student's } t=\frac{\sqrt{n-1}\cdot\overline{\Delta z}}{\sqrt{1/n\sum(\Delta z_{0i}-\overline{\Delta z})^2}}=-0.717$$

where n: number of samples : 10

Δz_{0i}: : shown in table 2

$\overline{\Delta z}$: arithmetic mean of Δz_{0i} : -0.322.

The random variable that corresponds with t, shows t-distribution with degree of freedom $n-1=9$. Then we have $\lambda=2.262$ from a statistic diagram, where λ satisfies

$$\Pr\{|X_9|>\lambda\}=\alpha=0.05$$

and X_9 is a function of t-distribution with degree of freedom 9. Therefore $|t|=0.717<\lambda=2.262$. So one can not reject the above hypothesis. In other words, the mean value may be zero.

According to this test, one may decide that the example in question has no systematic error with the confidence coefficient of 95%.

5. At last, one should estimate the variance of distribution of Δz_{0i}. Since the population mean of Δz_{0i} is fixed to be zero, the estimate of variance is calculated as follows.

$$\hat{\sigma}^2=1/(n-1)\sum(\Delta z_{0i}-\overline{\Delta z})^2=2.016$$
$$\hat{\sigma}=1.42.$$

λ that satisfies $\Pr\{|\Delta z|>\lambda\sigma\}=\alpha=0.05$, is derived from a statistic diagram as $\lambda=1.96, \therefore \lambda\hat{\sigma}=2.78$ (unit: meter). This is the absolute value of the standard error of z_{01}, z_{02} $\cdots\cdots$, and also must be that of z_1, z_2 $\cdots\cdots\cdots$ under the confidence coefficient of 95% in view of the above 2.

6. The error thus obtained is nothing but an accidental error defined by the writer. When this error does not meet the purpose of the investigation, the accuracy of the measuring method must be raised still more. But when this error is small enough to answer the purpose, the measurement is finished.

5. Acknowledgments

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An addition to the Mammalian fauna of the Japanese Miocene

By

Fuyuji TAKAI

(With Text-figs. 1a–c)

As Cenozoic formations in the Japanese Islands belong mostly to marine deposits, fossil occurrences of mammalians which have migrated from the Asiatic continent are quite restricted chronologically. Seven mammalian faunules have been discriminated by the writer in 1938, among which the Medial and Late Miocene faunules as well as the Pleistocene one are comparatively rich. These two, Hiramakian and Togarian, faunules must correspond respectively to the European Burdigalian and Vindobonian. The former consists of the following six species, namely, *Bunolophodon annectens*, *Anchitherium hypohippoides*, *Palaeotapirus yagii*, *Chilotherium pugnator*, *Brachyodus japonicus*, and *Amphitragulus minoensis* and the latter of *Sinanodelphis izumidaensis*, *Ontocetus oxymycterus*, *Idiocetus tsugarensis*, and *Desmostylus japonicus*.

A fossil suid here described under the new specific name of *Palaeochoerus japonicus* has been obtained from the Mimasaka coal-bearing beds at the Kitasaka lignite colliery of Yunogo-mura, Katsuta-gun, Okayama Prefecture. It is a fragmental right lower jaw with the first and second molars. According to K. SUYARI's geological survey the Tertiary formations of this area are divisible into the upper, Katsuta beds and the lower, Mimasaka coal-bearing beds. The former which yields several kinds of marine fossils, such as *Operculina complanata japonica*, *Vicarya callosa* and others are unconformably overlain by the Pleistocene Nihonbaru gravel beds. The stratigraphic relationship between the latter and the basement, composed of the Meso-Palaeozoic formations and igneous rocks (diorites and rhyolites), is unconformable.

The generic and specific identification of a suid with teeth, especially with lower teeth only, is a very difficult task for even keen specialists, but from its tooth-pattern it may be unquestionable that

the present material belongs to the family Suidae. G. G. SIMPSON has classified this family into twenty-two genera, among which the following fourteen occur in the Neogene deposits of Asia, namely,

Palaeochoerus POMEL, 1847	U. Mioc.–M. Plioc.
Chleuastochoerus PEARSON, 1928	L. Plioc.
Listriodon von MEYER, 1846	U. Mioc.–M. Plioc.
Conohyus PILGRIM, 1926	U. Mioc.–M. Plioc.
Tetraconodon FALCONER, 1868	L. Plioc.–Pleist.
Sivachoerus PILGRIM, 1926	M. Plioc.–Pleist.
Sanitherium von MEYER, 1866	L. Plioc.
Propotamochoerus PILGRIM, 1926	L.–M. Plioc.
Hyosus PILGRIM, 1926	M. Plioc.
Sivahyus PILGRIM, 1926	M. Plioc.
Hippohyus FALCONER and CAUTLEY, 1840–45	L. Plioc.–Pleist.
Dicoryphochoerus PILGRIM, 1926	L. Plioc.–Pleist.
Sus LINNAEUS, 1758	L. Plioc.–Recent
Lophochoerus PILGRIM, 1926	L. Plioc.

As stated above, the Katsuta beds, occupying unconformably upon the Mimasaka coal-bearing beds, are characterized by the occurrences of typical Miocene species of *Operculina complanata japonica* and *Vicarya callosa*. Therefore the geological age of the Mimasaka coal-bearing beds are considered to belong probably to Miocene instead of Pliocene in age. And also the stratigraphic relationship between the Katsuta and Mimasaka coal-bearing beds must be referrable to that between the Togarian and Hiramakian stages. Judging from this stratigraphic view-point the present specimen must be identified with any among *Palaeochoerus*, *Listriodon*, and *Conohyus*. On the other side the present specimen closely resembles those of *Palaeochoerus*, *Chleuastochoerus*, and *Dicoryphochoerus* from its dental characteristics. For the present moment it seems to be reasonable that the present material is referred to the genus *Palaeochoerus* on account of both its dental characteristics and stratigraphic view-points. The description of the new material is given below.

Palaeochoerus POMEL, 1847
Generic type, *Palaeochoerus typus* POMEL
Palaeochoerus japonicus TAKAI, n. sp.
Text-figs. 1a–c.

Material:—A fragmental right lower jaw with the first and second molars embedded in the Mimasaka coal-bearing beds at the Kitasaka lignite colliery of Yunogo-mura, Katsuta-gun, Okayama Prefecture; stored in the Museum of Geology and Paleontology, Faculty of Science, Tohoku University; Reg. No. 72697.

Description:—$\overline{M1}$ is broader in its posterior than in its anterior lobe. It is trapezoidal in outline and two-rooted. There are four main cusps of equal size, two median tubercles, and a posterior talon. A small median tubercle can be observed in the transverse valley, while a still smaller one juts out in front of the median part of the anterior lobe. The posterior talon is located just behind the posterior lobe. The cingulum is present, but weak. $\overline{M2}$ possesses the similar tooth-pattern of $\overline{M1}$, but is a little larger in size.

Fig. 1. *Palaeochoerus japonicus* TAKAI, n. sp. ×2.
a. crown view of right first and second molar
b. lingual view of the same.
c. buccal view of the same.

Measurements in mm.

$\overline{\text{M1}}$ { antero-postero............14.8
transverse 12.1

$\overline{\text{M2}}$ { antero-postero............18.9
transverse 13.8

Remarks:—Its reference to the genus *Palaeochoerus* is tentative. The following eleven species have ever been described under this genus, namely,

Palaeochoerus typus POMEL, 1847 (Olig.–L. Mioc., Europe)
P. simplex (GERVAIS), 1852 (M. Mioc., France)
P. major POMEL, 1847 (L. Mioc., France)
P. waterhousei POMEL, 1853 (L. Mioc., France)
P. meissneri von MEYER, 1850 (M. Mioc., Europe)
P. choeroides (POMEL), 1848 (M. Mioc., Europe)
P. helveticus (PICTET and HUMBERT), 1869 (Eoc., Switzerland)
P. aurelianensis STEHLIN, 1899 (Mioc., France)
P. pascoei PILGRIM, 1926 (The Murree zone, Burma)
P. perimensis (LYDEKKER), 1887 (The Nagri zone, India)
P. lahirii PILGRIM, 1926 (The Kamlial zone, India)

Among the European species *Palaeochoerus aurelianensis* has the most intimate akinship with the present material, but is clearly discriminated by its much blunt cusps and its still bigger size. It is a very difficult task to compare the present material with the Indian species being described insufficiently by several isolated teeth. The exact phyletic relationship between the Japanese and Indian on one side and the European species on the other shall be decided by future occurrences of more complete specimens.

This paper embodies the result of one of the studies under the Science Research Fund from the Department of Education. The writer wishes to express his many thanks to the Department as well as to Profs. Hisakatsu YABE and Shoshiro HANZAWA for placing this new material at his disposal. Acknowledges and thanks are also due to Messrs. Yoriyuki TAKO and Chotaro EGASHIRA both of the Kitasaka lignite colliery for their generous assistances offered in field. To Prof. Teiichi KOBAYASHI of our Institute he is also indebted for his encouragement and advice.

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On the Occurrence of Warwickite $(Mg, Fe)_3TiB_2O_8$ at Hol Kol, Korea: a Study of Boron Metasomatism

By

Takeo WATANABE

(With Plates IV–V)

Abstract

Warwickite, a titano-borate of magnesium and iron, is a very rare mineral which has hitherto been found only in crystalline limestone near the town of Warwick, Orange County, New York. The second new occurrence of this mineral is at the Hol Kol mine, Suan, Korea. It occurs as a boron metasomatic product in kotoite-marble bordering gold-copper-bearing diopside-clinohumite skarn masses. The associated minerals are ludwigite, spinel, forsterite, clinohumite, pyrrhotite and calcite.

Introduction

Warwickite was first discovered by SHEPARD[7,8] (1832–38) in the crystalline limestone at Edenville near Warwick, Orange County, New York State, but detailed and chemical descriptions were given later by HUNT[3,4] (1851), SMITH[9] (1874) and BRADLEY[1] (1909) and its optical properties were determined by LACROIX[5] (1886). The paragenesis of warwickite was briefly described by WHITLOCK[16] (1903). The crystal structure was determined by TAKEUCHI et al.[10] (1950). However, no detailed paragenetic study on this mineral has recently been done. In the course of his study on boron minerals the writer discovered the second known occurrence of this mineral at Hol Kol, Suan, Korea and studied its paragenetic relations to other titanium-bearing minerals.

It is the purpose of the present paper to describe the new occurrence of warwickite at Hol Kol and to discuss the origin of titanium-bearing minerals in dolomite and magnesian carbonate rocks associated with the gold-copper-bismuth-bearing skarn ore.

Occurrence

The new occurrence of warwickite is at the Hol Kol mine, Suan County, Hwanghai-do, North Korea. The mineral occurs in a sheath of kotoite-marble which encloses the gold-bearing diopside skarn pipe of the North Ore Body of this mine.

The kotoite-marble consisting essentially of kotoite, $Mg_3B_2O_6$, and calcite, is a remarkable rock formed from dolomite by boron metasomatism in the contact zone of the Suan granite. The detailed descriptions of the paragenesis of borates at Hol Kol were given by the writer in 1939[13] and 1943[14]. At that time the writer recorded two unknown minerals found in the kotoite-marble. Last year one of the unknown minerals was identified by the writer as a new magnesium borate mineral, $Mg_2B_2O_5$, and was named suanite[15], and the other brown mineral was also determined as warwickite. The warwickite is sporadically found in the white kotoite-marble and is especially richly present in brown bands of the kotoite-marble as shown in Pl. IV, Figs. 1–2, and Pl. V, Fig. 3.

It is very interesting to observe that this brown band in the kotoite-marble connects with the black band in dolomite-marble immediately adjacent to the kotoite-marble. These dark bands in dolomite-marble, consisting mainly of forsterite, spinel, geikielite (or picroilmenite), pyrrhotite, and calcite, may represent original impure siliceous and aluminous layers in dolomite. Although geikielite had been known as a very rare mineral, it was first found in place by J. MURDOCH[6] (1949) in California. This geikielite occurs also sparingly in a spinel rich zone of the brucite-marble derived probably from a dolomitic rock. It is associated with forsterite, diopside, spinel, brucite, calcite, pyrite, and pyrrhotite. Murdoch considered that the geikielite was a recrystallized product and of metamorphic origin. It seems also to the writer that the titanium content was originally high in the aluminous layers in the dolomite found at Hol Kol. Isolated minute grains of geikielite from Hol Kol also show their beautiful crystal faces. Its habit is tabular parallel to base. The color of the geikielite from Hol Kol is a little darker than that of California when compared under the microscope. It is weakly pleochroic with ε, pinkish red, and ω, reddish brown. Under the ore-microscope the properties of the geikielite are very similar to those of ilmenite except that the former shows distinct internal reflection with pinkish color. More detailed descriptions of the geikielite together with its paragenesis will be given in a separate paper.

At the boundary between white dolomite-marble and yellowish kotoite-marble, color of the black band changes to brown as explained in Pl. IV, Figs. 1–2. It is seen under the microscope that

the brown bands in the kotoite-marble contain abundant grains of spinel, forsterite and radiated fibrous aggregates of warwickite with subordinate ludwigite. On the other hand the black bands in the dolomite, contain geikielite, magnetite and pyrrhotite abundantly and they connect with ludwigite-bearing warwickite-bands in the kotoite-marble.

Physical and optical properties

The warwickite from Hol Kol is dark brown in color. Its hardness is 3.5 and its specific gravity measured by suspension method is 3.3. Under the microscope it is light yellow to reddish brown in color. It occurs in prismatic to acicular crystal with rhombic section, which often forms a felt of fibres as shown in Plate V, Fig. 2. Its prisms and fibres are parallel to c and X; therefore, the character of zone is negative. Indices of refraction were determined by immersion method as shown in Table 1.

Table 1. Optical properties of warwickites

1. By T. Watanabe   2. After Larsen & Berman

		1	2
Material		Hol Kol	Edenville
α		1.772	1.808
β		n.d.	1.810
γ		1.792	1.830
$\gamma-\alpha$		0.020	0.022
2V		n.d.	(+) 59°
Opt. orientation		X=C (Prism.)	X=C (Prism.) Z=a
Pleochroism	X	brown	clear yellowish brown
	Y	light brown	reddish brown
	Z	light brown	cinnamon brown
Absorption		X>Y=Z	X>Y>Z

The accurate determination of its indices was very difficult on account of small size and slenderness of its crystals. As shown in Table 1 refractive indices of the warwickite from Hol Kol are lower than those of the original one probably due to the lower content of iron. It is also interesting to note here that the associated minerals such as spinel and geikielite from Hol Kol, are generally lighter in color

than those from Warwick. Other optical properties are also given in the Table 1.

Chemical properties

The mineral is soluble in warm hydrochloric acid and readily soluble in H_2SO_4. Although a great quantity of the kotoite-marble was dissolved in dilute HCl, the writer could not obtain enough material for chemical analysis. However, qualitative analysis showed the presence of boron and titanium in it.

X-ray powder photograph

The identity of the Hol Kol mineral and the original warwickite on the basis of X-ray powder methods was confirmed through the kindness of Dr. TAKEUCHI of the Mineralogical Institute and Mr. OSSAKA of the Earthquake Research Institute, University of Tokyo. The powder photographs taken by OSSAKA are reproduced in Pl. V, Fig. 1. The result of measurements of X-ray powder photographs is also given in Table 2.

Origin of warwickite and boron metasomatism

Warwickite and geikielite, two titanium minerals have long been known as very rare minerals and their genetical relationship has little been discussed. The present work on the warwickite from Hol Kol may present some interesting geochemical data for boron-metasomatism.

It has generally been considered that titanium content in such carbonate rocks as limestone and dolomite is comparatively low. However, a small amount of titanium is always found in dolomite or limestone. In that case the titanium appears to be associated with clayey materials which are contained as impurities in those carbonate rocks. (GOLDSCHMIDT[2] (1954)).

In 1949 Murdoch recorded a new occurrence of geikielite in brucite-marble from California, which was clearly of contact metamorphic origin. According to him, the geikielite occurs in spinel rich layers in the brucite-marble as in the case of Hol Kol dolomite. It is also of interest that picroilmenite and ilmenite have often been recorded in crystalline limestone from Edenville, New York, where

warwickite was first found. It seems, thus, probable that the occurrence of geikielite or picroilmenite in contact metamorphosed dolomite and magnesian limestone may be attributed to titanium content in the original rocks. As shown in Pl. IV, Figs. 1 and 2, the gold-bearing diopside-clinohumite ore skarn is bordered by a narrow rim consisting

Table 2. X-ray Powder Data for Warwickites

1, 2, 3; Chromium radiation, 28.5 mm. diameter camera.
4 ; Iron radiation (λ=1.937A). After R. M. THOMPSON & J. A. GOWER[11] (1954).

1		2		3		4	
Hol Kol, Suan, Korea		Edenville, New York		Synthetic ($3MgOB_2O_3TiO_2$)		Edenville New York	
I	d	I	d	I	d	I	d
		4	6.68	1	6.48	4	6.55
		1	4.60				
1	4.17	4	4.25	1	4.19	4	4.20
						2	3.73
3	2.92	4	2.99	3	2.97	3	2.97
3	2.84	2	2.86	2	2.82	½	2.84
				1	2.73	5	2.75
10	2.56	10	2.58	10	2.57	10	2.57
2	2.50	1	2.50	2	2.48	1	2.48
1	2.47	1	2.45				
		1	2.31			1	2.24
		1	2.19	1	2.17	½	2.20
				2	2.11	1	2.13
4	2.07	4	2.10	4	2.08	2	2.08
		1	2.02	1	2.00		
5	1.97	3	1.98	3	1.98	3	1.979
						1	1.867
1	1.83	2	1.85	1	1.83	½	1.838
		1	1.82			½	1.813
3	1.78	2	1.79	3	1.79	1	1.791
		1	1.77	1	1.76		
		1	1.66	1	1.72	3	1.721
						1	1.632
4	1.59	5	1.60	4	1.60	2	1.595
4	1.54	4	1.55	4	1.54	1	1.553
4	1.49	5	1.50	4	1.50	3	1.501
		2	1.43	1	1.46	1	1.471
		2	1.40	1	1.42		
3	1.37	3	1.38	2	1.38	2	1.380
						1	1.338
2	1.32	1	1.32	2	1.32	1	1.327
2	1.28	2	1.29	2	1.29	½	1.284
2	1.26	3	1.27	2	1.28	1	1.272
				3	1.24	½	1.243
				2	1.22	½	1.226
						1	1.184
						2	1.152

of ludwigite, fluoborite, szaibelyite, clinohumite, and calcite. The ore skarn as a whole is theathed by a somewhat wider zone of the kotoite-marble. The original texture of dolomite has been well pre-

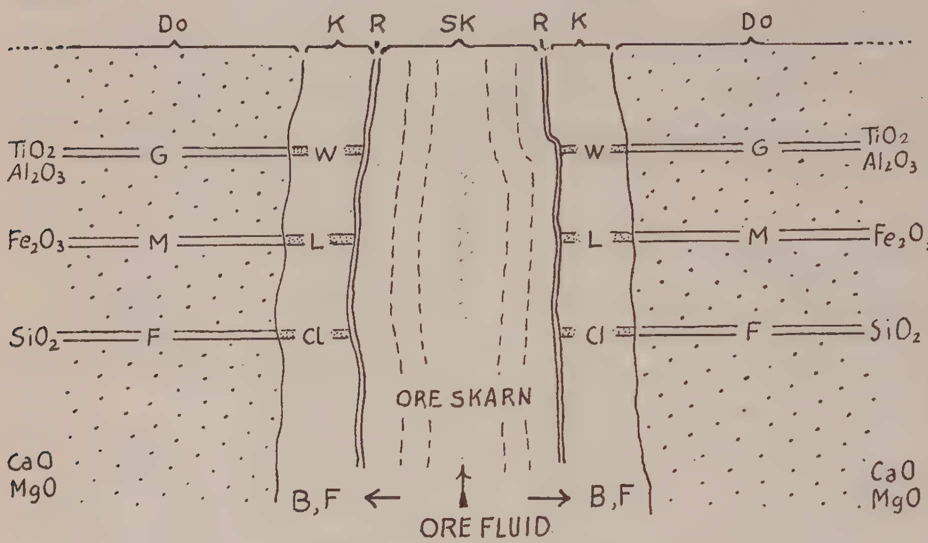

Fig. 1. Diagrammatic section of ore skarn pipe of the North Ore Body, Hol Kol Mine, Suan, Korea.

SK: gold-copper-bearing diopside-clinohumite skarn;
R: reaction-rim consisting of ludwigite, fluoborite, szaibelyite, clinohumite, and calcite;
K: kotoite-marble, a boron metasomatic product;
Do: dolomite-marble.

Horizontal layers represent impure layers in dolomite. The diagram showing the paragenetic changes of the impure layers due to boron metasomatism.

G: geikielite-spinel layer in dolomite-marble;
W: warwickite layer in kotoite-marble:
M: magnetite-pyrrhotite layer in dolomite-marble;
L: ludwigite layer in kotoite-marble;
F: forsterite-calcite layer in dolomite-marble;
Cl: forsterite-clinohumite layer formed by fluorine metasomatism.

served in this kotoite-marble zone. Then, the kotoite-marble gives way sharply to a normal recrystallized dolomite-marble. At the contact between these marbles the geikielite-bearing dark layers are usually well traced from the dolomite side into the kotoite-marble. The change of color of the dark layers of the marbles at the contact is due to the disappearance of geikielite and the appearance of warwickite. Thus,

it may be concluded that the warwickite has newly been developed from the geikielite-bearing dolomite-marble by boron metasomatism at the time of kotoitization, according to the hypothetical equation:

$$\underset{\text{dolomite}}{2CaMg(CO_3)_2} + \underset{\text{geikielite}}{MgOTiO_2} + B_2O_3 = \underset{\text{calcite}}{2CaCO_3} + \underset{\text{warwickite}}{3MgOTiO_2B_2O_3} + 2CO_2$$

It is also noted that the law of equal volume has been kept in the formation of the kotoite-marble by boron-metasomatism as seen in Plate IV, Figs. 1–2. The general paragenetic relations of magnesium-borate minerals observed in the Hol Kol area are shown in Fig. 1.

The development of the ore skarn and the kotoite-marble at Hol Kol must be attributed to ore-bearing solutions ascending from below along the contact zone of the Suan granite. The zoning of skarn minerals and borate minerals surrounding ore-pipe of the Hol Kol mine may be also interpreted as a diffusion effect like the Liesegang phenomenon, because the original texture due to the impure layers in the dolomite has well been preserved.

The similar association of multizoned skarns with magnesium-borate minerals in Skye, has already been described by C. E. TILLEY[12], although warwickite has not been discovered there.

It seems, however, highly probable that a similar paragenesis as at Hol Kol will be found in other places where magnesium borate-minerals occur in the dolomite-marble near a granitic intrusion.

Acknowledgements

The writer wishes to express his thanks to Prof. T. ITO for his kind advice, to Dr. Y. TAKEUCHI, Mr. R. MORIMOTO and Mr. J. OSSAKA for their X-ray work and to the late Mr. Taira NAKANO for his kind help in laboratory work. Grateful acknowledgements are also made to Dr. W. F. FOSHAG of U. S. Nat. Museum, Prof. H. H. HESS of the Princeton University, Prof. H. WINCHELL of the Yale University and Prof. J. MURDOCH of California University at Los Angeles for samples of warwickite and geikielite which were presented to the writer for the comparison study.

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Plate IV.

Explanation of Plate IV.

Fig. 1–2. Polished surface of zoned skarn from the North Ore Body, Hol Kol, showing paragenetic relation of Ti-bearing impure layers at the contact between dolomite-marble (Do) and kotoite-marble (K). W, warwickite layer in kotoite-marble: G, geikielite-spinel layer in dolomite-marble. R, reaction rim between diopside (Di) clinohumite (Cl) ore skarn and kotoite-marble. Black fine-needles in kotoite-marble are ludwigite. Black parts in ore skarn consist of gold-copper-bismuth bearing sulphide ore minerals.

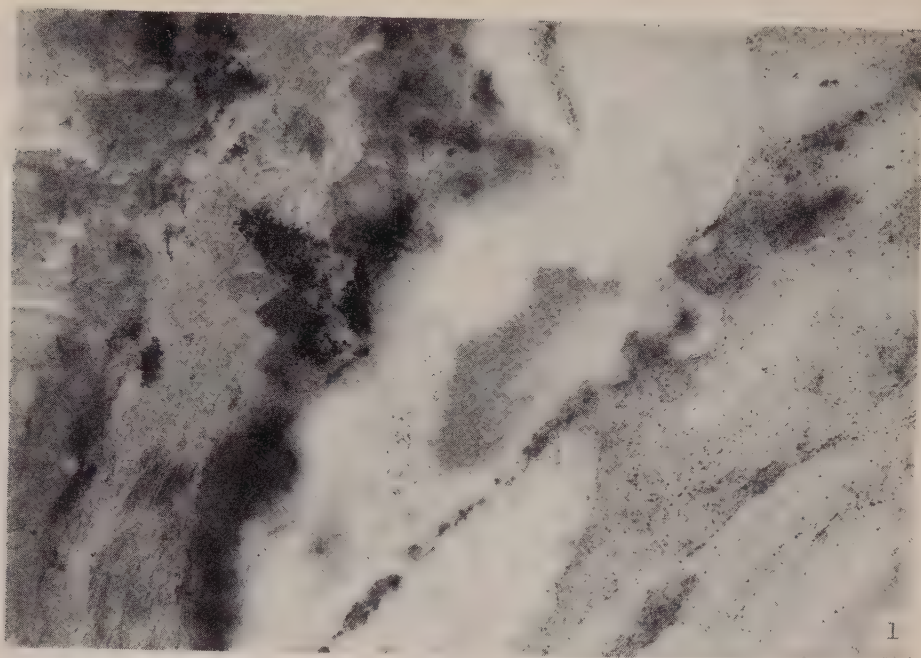

Fig. 1

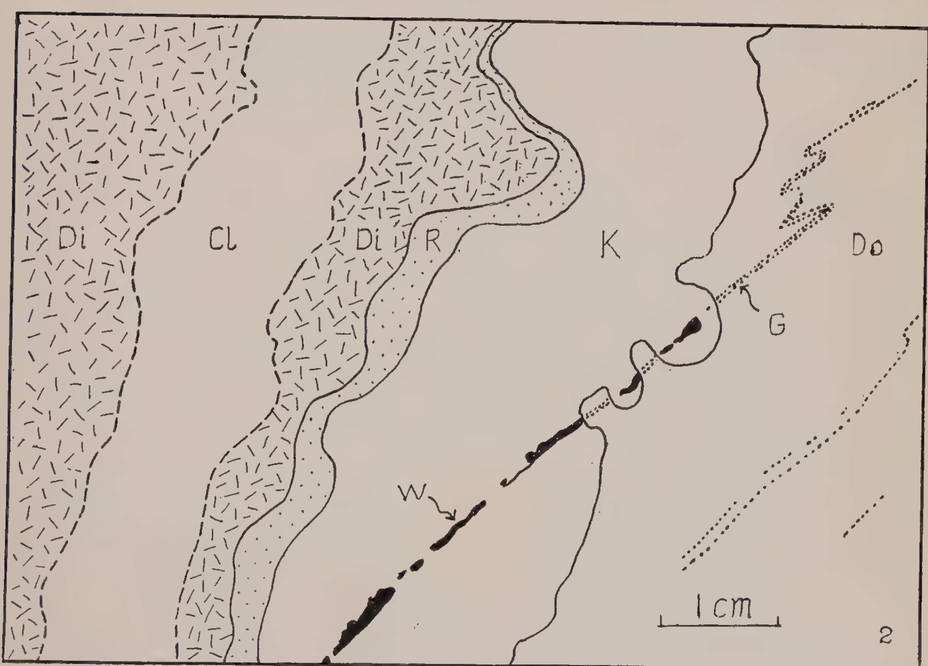

Fig. 2

Plate V.

Explanation of Plate V.

Fig. 1. X-ray powder photographs of (1) warwickite (Hol Kol), (2) warwickite (Edenville), (3) Synthetic magnesio-warwickite ($Mg_3TiB_2O_8$). Cr radiation: camera 28.5 mm diameter.

Fig. 2. Photomicrograph of thin section of warwickite-spinel layer in kotoite-marble. W, warwickite; L, ludwigite; Sp, spinel; F, forsterite; C, calcite. ×70.

Fig. 3. Polished surface of warwickite-bearing kotoite-marble (K). Black layers are very rich in warwickite (W), spinel, forsterite and ludwigite; Kotoite-marble with impure layers cut by szaibelyite veins (Sz).

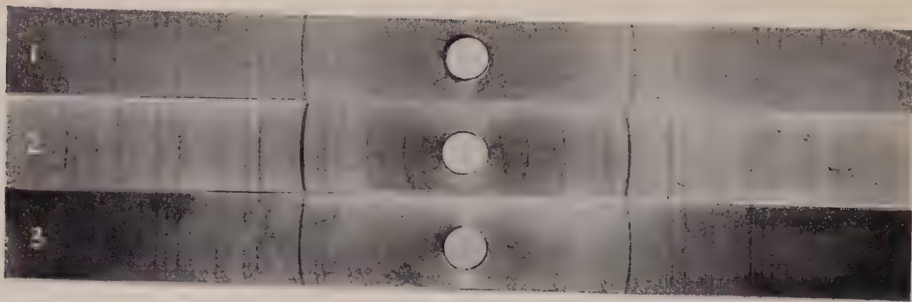

Fig. 1

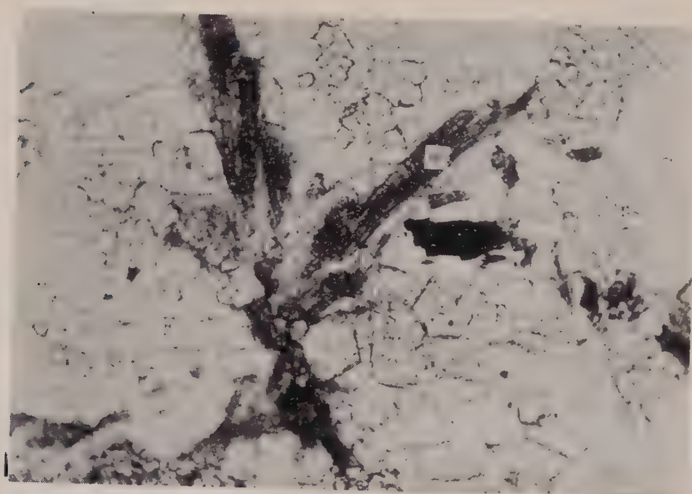

Fig. 2

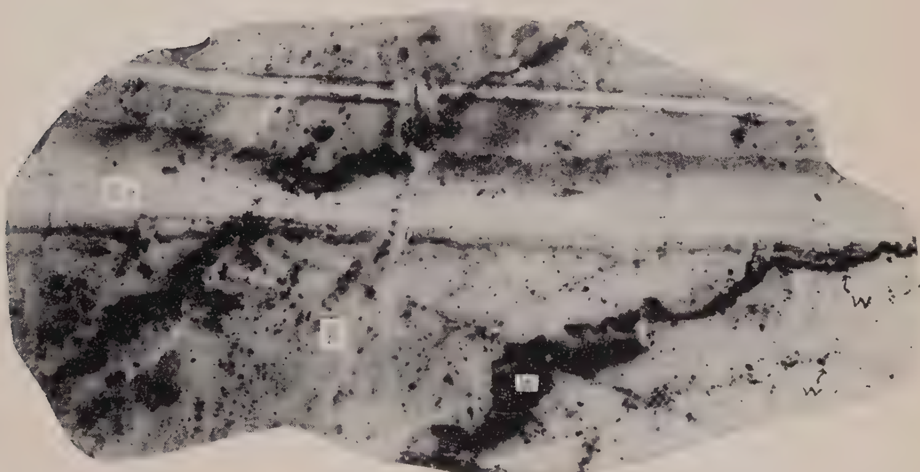

Fig. 3

On the Chemical Composition of Lavas of Nyohô-Akanagi Volcano, Nikkô.

By

Masao YAMASAKI

Abstract

Chemistry of the rock series in Nyohô-Akanagi volcano is controlled by fractional crystallization as well as by contamination by argillaceous sediments underlying the volcano. It is demonstrated that through fractional crystallization the ratio K_2O/Na_2O in the lavas rises slowly with SiO_2, while it rises markedly through contamination because the argillaceous sediments have a very high K_2O/Na_2O ratio. The lavas erupted later in the history of the activity of the volcano represent generally more advanced stages of crystallization and also of contamination.

Introduction

Nyohô-Akanagi (2463.5 m) is a strato-volcano lying about 130 km north of Tokyo. The center of the volcano is located about 8 km northwest of the city of Nikkô, Tochigi Pref. Among the members of the Nikkô volcano group which belongs to the Nasu volcanic zone of northeastern Japan, Nyohô-Akanagi is the oldest one. It is a dissected volcano without any record of activity during the historic time. Its age is probably Pleistocene.

The geology of the Nikkô district was described by SAITÔ (1899), TSUBOI and SUGI (1926) and IWAO (1943), but only brief descriptions were given on the volcano. In the present paper the writer intends to give a summary of the results of a chemical study of the Nyohô-Akanagi volcano; a more detailed description of the geology and petrography will be given on a later occasion.

The writer wishes to express his sincere thanks to Dr. Seitarô TSUBOI, Dr. Hisashi KUNO, Dr. Akiho MIYASHIRO and other members of the Geological Institute, Tokyo University, for their kind advice and criticism during this work. He is also much obliged to Prof. Shûichi IWAO for the permission to use unpublished data on the geology of the district and to Mr. Jun ITÔ for his kind help in chemical analysis. Messrs. Yoshiharu MATSUMURA, Shichirô NAKAMURA and other

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Outline of Geology

The basement of Nyohô-Akanagi consists of unclassified Palaeozoic strata intruded by granitic rocks, and a complex of rhyolitic rocks, including welded tuff, of middle Tertiary age. This basement forms a broad ridge which extends from east to west and lies between two rivers, Kinu on the north and Daiya on the south. The crest of the ridge is generally about 600 m high above the bottom of the two rivers.

The center of eruption of the volcano was situated near the crest of the ridge, and the materials piled up around this center reached a thickness of more than 1200 m. The crater which existed at the top of the cone was breached on southern side by the River Inari, resulting in a steep-sided, horseshoe-shaped depression with a diameter of about 2 km. Where the river crosses the crater wall, it forms a gorge, named Unryû, about 600 m deep. Nyohô and Akanagi are the names of the western and eastern peaks respectively on the northern crater wall.

The sequence of eruption of the lavas is generally from basic to acidic, although acidic and basic lavas were erupted alternately at some stage of the history of activity.

In Table 1 are shown the chemical compositions of the main lavas of the volcano. (No. 1 to No. 9) They are arranged in the order of eruption as determined chiefly on the southwestern wall of the gorge of Unryû.

Description of Analysed Samples

Among the earlier lavas, basic pyroxene andesites with or without olivine phenocrysts are dominant. The specimens Nos. 1 and 2 are the examples of the lavas of this stage:

No. 1: Augite-hypersthene-olivine andesite (MY53101304). Loc. Shimotakizawa, Kuriyama at the northern foot of the volcano. A lava which flowed down northward.

No. 2: Hypersthene-augite andesite (MY5210024). Loc. Tainai-Baku at the bottom of the gorge of Unryû.

Table 1. Chemical Compositions and Norms of Lavas of Nyohô-Akanagi Analyst, M. YAMASAKI.

	1	2	3	4	5	6	7	8	9	A	B
SiO_2	52.29	55.91	64.96	53.77	62.51	62.00	64.70	62.56	56.20	48.73	71.01
Al_2O_3	17.98	18.45	15.69	19.59	16.99	16.49	15.68	15.95	15.19	16.53	13.85
TiO_2	0.81	0.93	0.44	0.55	0.67	0.11	0.69	0.69	1.15	0.63	0.65
Fe_2O_3	3.44	3.59	1.78	4.01	2.07	4.21	2.74	2.69	4.38	3.37	1.11
FeO	6.46	4.52	3.29	5.54	3.57	3.73	2.87	3.09	4.54	8.44	3.24
MnO	0.20	0.13	0.12	0.14	0.10	0.13	0.14	0.11	0.20	0.29	0.03
MgO	5.07	2.96	1.35	3.31	1.86	2.22	2.61	2.61	5.34	8.24	2.50
CaO	10.05	8.38	4.83	8.80	5.88	5.56	5.69	5.54	6.80	12.25	0.32
Na_2O	1.86	3.13	4.16	2.45	3.48	3.16	3.19	3.43	2.62	1.21	1.28
K_2O	0.33	0.61	1.27	0.89	1.58	1.44	1.80	1.86	1.63	0.23	3.20
$H_2O(+)$	0.70	1.01	1.09	1.03	0.82	0.32	0.35	0.55	1.25	—	1.74
$H_2O(-)$	0.38	0.63	0.52	0.33	0.37	0.20	—	0.21	1.11	—	0.49
P_2O_5	—	0.14	0.24	0.11	0.21	0.22	—	0.20	0.21	0.10	0.08
Total	99.57	100.39	99.74	100.52	100.11	100.21	100.46	99.49	100.62	100.02	99.50
Q	9.65	12.88	22.79	10.48	20.02	21.92	23.31	19.89	12.75		C=n.d.
or	1.95	3.62	7.51	5.29	9.35	8.52	10.63	11.02	9.63		
ab	15.73	26.47	35.14	20.71	29.37	26.74	27.00	28.97	22.16		
an	39.75	34.49	20.38	39.83	26.09	26.19	23.15	22.64	24.86		
C	—	—	—	—	—	0.12	—	—	—		
wo	4.22	2.60	0.89	1.31	0.77	—	2.13	1.53	3.18		
en	12.62	7.37	3.36	8.24	4.63	5.53	6.49	6.49	13.29		
fs	8.07	4.75	4.06	6.21	3.92	3.42	2.14	2.52	3.19		
mt	4.98	5.21	2.60	5.81	3.01	6.11	3.98	3.91	6.34		
il	1.54	1.76	0.84	1.05	1.28	0.21	1.31	1.31	2.18		
ap	—	0.34	0.57	0.27	0.51	0.54	—	0.47	0.50		
Or (in mol. %)	3	5	12	8	14	13	17	17	12		
Ab (in mol. %)	29	43	57	33	47	45	46	48	28		
An (in mol. %)	68	52	31	59	39	42	37	35	60		
Wo (in mol. %)	16	18	11	8	8	—	19	14	15		
En (in mol. %)	57	58	46	59	56	68	65	67	72		
Fs (in mol. %)	27	24	43	33	36	32	16	19	13		

Table 2. Modes of Analysed Samples (in volume %).

	1	2	3	4	5	6	7	8	9
Phenocrysts: Quartz	—	—	—	—	—	—	—	1.0	—
Phenocrysts: Plagioclase	18.1	29.6	0.8	28.1	18.1	2.7	25.1	23.2	6.5
Phenocrysts: Augite	1.3	4.9	tr.	1.2	1.8	0.2	2.3	1.6	0.7
Phenocrysts: Hypersthene	3.8	3.3	0.4	7.5	1.0	0.2	2.5	2.2	1.3
Phenocrysts: Olivine	3.4	tr.	—	tr.	—	—	—	—	—
Phenocrysts: Hornblende	—	—	—	—	—	—	1.2	2.2	—
Phenocrysts: Magnetite	—	1.7	tr.	2.2	1.5	0.5	1.6	1.9	—
Groundmass	73.4	60.5	98.8	61.0	77.6	96.4	67.3	67.9	91.5

At an earlier stage, however, eruption of some acidic andesites also took place as shown by the following example;

No. 3: Hypersthene-augite-bearing andesite (MY5210028). Loc. 50 m above the bottom of the gorge. An almost non-porphyritic lava extending southward as far as the present city of Nikkô.

On the western wall of the crater, the lava No. 3 is overlain successively by the following three lavas.

No. 4: Olivine-bearing augite-hypersthene andesite (MY5210029). Loc. 150 m above the bottom of the gorge. A porphyritic lava which flowed down chiefly southward.

No. 5: Hypersthene-augite andesite (MY52100210). Loc. 300 m above the bottom of the gorge. Several layers of lavas exposed on the eastern cliff along the River Inari consist of the same rock. These lavas flowed down chiefly to the east and south.

No. 6: Augite-hypersthene andesite (MY5209221). Loc. South of Fujimi pass. The western half of the slope of Nyohô is veneered by this lava. Distinct flow banding is characteristic.

The chief groundmass constituents of the lavas Nos. 1–6 are plagioclase, monoclinic pyroxene, iron ore, tridymite and interstitial glass. In the groundmass of Nos. 5 and 6, phlogopite and pargasite occur, often forming minute patchy aggregates with quartz.

The activity of the volcano closed with the extrusion of hornblende andesite and dacite of Nos. 7 and 8.

No. 7: Green hornblende-augite-hypersthene andesite (MY53100801). Loc. Uehara, Kuriyama. The rock was collected from one of the essential fragments in the pyroclastic flow (nuée ardente) which spreads northeastward and forms the flat terrace of Uehara on the northern slope of the volcano. Besides these fragments of compact andesite, the deposit includes accidental fragments of Palaeozoic rocks in a matrix of volcanic ash.

No. 8: Augite-hypersthene-hornblende dacite (MY5209302A). Loc. Near the summit of Nyohô. This lava was erupted at the end of the activity of the volcano. The lava was so viscous that it piled up only around the crater. Phenocrysts of hornblende are completely altered to pyroxene opacite. Sporadic quartz crystals may be xenocrysts derived from Palaeozoic siliceous sediments. The abundance of cognate inclusions is a distinctive character of the lava.

No. 9: Augite-hypersthene porphyrite (MY5209302B). A cognate inclusion in No. 8.

The groundmass of the last three rocks consists of plagioclase, hypersthene, augite, iron ore and tridymite. In the groundmass of Nos. 8 and 9, phlogopite and pargasite also occur.

The writer agrees to KUNO's (1950) view that cognate inclusions represent aggregate of crystals separated from magma in order to supply the heat necessary to dissolve foreign materials captured by the magma. Some of the above cognate inclusions in the lava No. 8

contain in their core fragments of microdioritic rocks. These rocks are composed chiefly of hypersthene, augite and plagioclase, and seem to have recrystallized from some Palaeozoic sediment. The presence of plagioclase with a zone of dust inclusions both in the cognate inclusions and in the host lava suggests that the lava was contaminated by foreign materials as was discussed by KUNO (1950). Similar cognate inclusions or basic inclusions are reported from other volcanoes of the world; for example, Lassen Peak, California, and Crater Lake, Oregon (WILLIAMS, 1932 and 1942).

K_2O/Na_2O Ratios of Lavas

Fig. 1 shows the relation between K_2O/Na_2O (in wt.) and SiO_2 in lavas of Nyohô-Akanagi. The numbers for the points in the figure refer to those in Table 1. Considering the field relations of the rocks, curves are drawn to show the K_2O/Na_2O variation for each of the three groups of rocks: Nos. 1, 2 and 3, Nos. 4, 5 and 6, and Nos. 7 and 8.

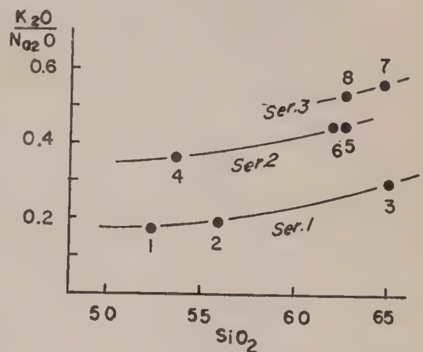

Fig. 2.

It is generally believed that calcic plagioclase, pyroxenes and olivine are the minerals to separate in an earlier stage of the crystallization of basaltic magma. Since these minerals are poor in alkalies, their separation will cause little change of the value of K_2O/Na_2O in the residual magma. As the crystallization proceeds, the composition of plagioclase separating from the magma becomes successively more sodic, and the value of K_2O/Na_2O in the residual magma will gradually increase. On the other hand, the SiO_2 content of the residual magma will increase continuously. Therefore the curve showing the relation between K_2O/Na_2O and SiO_2 will show only little inclination in the earlier stage of differentiation but will turn upward gradually in the later stage.

Thus the individual curves in Fig. 1 may be taken as showing the variation of the ratio K_2O/Na_2O due to the fractional crystallization of magmas. As the three groups of lavas follow independent courses

of fractionation as shown in the figure, they are called here Series 1, 2 and 3 respectively.

The stepwise increase of the ratio K_2O/Na_2O from Series 1 to 2 and also from Series 2 to 3 is not likely to be due to fractional crystallization; an effect of contamination of the magmas by the wall rocks of the reservoir should be considered.

Xenoliths are not uncommon in the lavas of Nyohô-Akanagi and many accidental fragments are also found among the ejecta. Bulk of them appears to have been derived from the Palaeozoic strata widely exposed in the surrounding area. This formation is made up largely of clay-slate and partly of chert and schalstein. The argillaceous sediment is most likely the material that was incorporated by the magma of the volcano and affected the K_2O/Na_2O ratio of the rock series.

Variation Diagram

In Fig. 2 are shown the variation curves for the oxides of the three rocks belonging to Series 1. These curves may be taken as the standard representing the oxide variation of the rocks derived from an original magma chiefly by fractional crystallization without appreciable contamination.

The curves, when extrapolated, pass the points for the oxides of the original magma of the pigeonitic rock series of Izu and Hakone as given in column A, Table 1 (KUNO, 1953). This fact suggests that the original magma of Nyohô-Akanagi had a composition very close to that of the pigeonitic rock series given by KUNO.

The composition of the lavas of Series 2 and 3 are shown by points in the same diagram. Some of the points show marked departure from the standard curves.

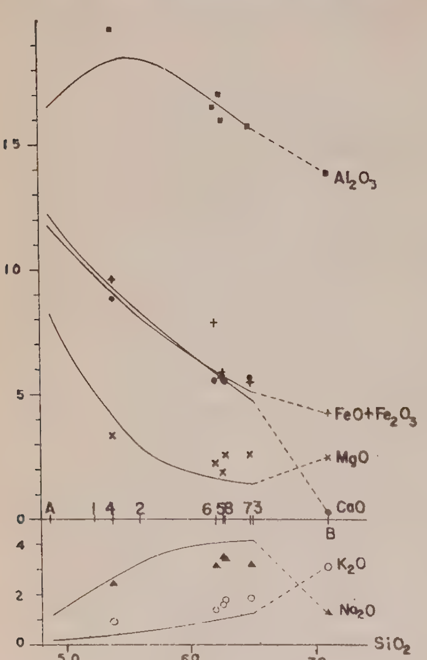

Fig. 2.

In column B, Table 1 is given the average composition of twelve specimens of argillaceous sediments collected from the Palaeozoic and Mesozoic formations exposed in the vicinity of Akagi volcano lying about 40 km southwest of Nikkô (ÔTA, 1952). It is assumed that the materials which were incorporated in the magma of Nyohô-Akanagi have similar compositions. This average composition is also plotted in Fig. 2. The point for each oxide is connected with the end of the corresponding curve by a broken line.

As was discussed by BOWEN, argillaceous sediments will be dissolved and incorporated in basaltic magma, for they consist of minerals corresponding to the later member of the reaction series (BOWEN, 1928). Therefore any rocks affected by contamination would have oxide composition represented by points lying on straight lines which connect the points for the argillaceous sediments and some points on the standard variation curves, if the effect of crystallization attending the contamination can be ignored. When the curves for CaO and $FeO+Fe_2O_3$ are extrapolated, they pass near the points for the corresponding oxides of the argillaceous sediments. So the curves nearly coincide with the straight lines in the addition-subtraction diagram of the corresponding oxides. Therefore the effect of the contamination on these oxides can hardly be distinguished from the variation caused by fractional crystallization.

On the other hand, the points for Na_2O and K_2O of the sediments lie below and above the extension of the corresponding curves respectively. Therefore these oxides can be used as a clue to distinguish the effect of the contamination from that of the fractional crystallization. All the points for K_2O and Na_2O of the lavas of Series 2 and 3 lie above and below the corresponding curves of Series 1 respectively. This fact strongly indicates that these lavas are contaminated by the sediments.

It is therefore expected that the effect of contamination is more clearly recognized if the ratio K_2O/Na_2O is plotted against SiO_2. The value of K_2O/Na_2O in the average composition of the sediments is about 2.5, while that of the lavas of Series 1 ranges from 0.18 to 0.31. The value rises abruptly from Series 1 to Series 2 and again from Series 2 to Series 3. In Series 2 it ranges from 0.36 to 0.45 and in Series 3 from 0.54 to 0.56.

Most of the argillaceous sediments are richer in K_2O than in Na_2O (RANKAMA and SAHAMA, 1948. See page 222). In his study of the

contamination process of gabbroic magma by metamorphosed argillaceous sediments in Aberdeenshire, READ (1923) described several examples of enrichment of K_2O in contaminated rocks and stated: "Potash in the majority of cases is increased; soda is rather indefinite, but seems upon the whole to increase; soda is always greater than potash in the initial magmas, but the disparty is either less marked, or potash is in excess of soda in the contaminated magmas."

In Fig. 2, the points for MgO of Series 2 and 3 lie above the corresponding curve for Series 1. The enstatite component in normative pyroxene increases from Series 1 to Series 3. This fact may also be explained as due to contamination.

Conclusion

During the activity of Nyohô-Akanagi, the character of the magma changed gradually. Thus, in the early stage, basic pyroxene andesites were erupted dominantly, while in the later stage acidic pyroxene andesites became more predominant among the erupted materials. The last activity of the volcano is the eruption of pyroclastic flow, succeeded by extrusion of viscous lava rich in cognate inclusions. The rocks of this stage are acidic andesite and dacite containing hornblende phenocrysts. A similar trend in the variation of the magma is also observed in the history of Nantai volcano adjacent to the southwest of Nyohô-Akanagi.

The collapse of caldera which often follows the eruption of pyroclastic flow did not take place either in Nyohô-Akanagi or in Nantai.

During the magmatic evolution of Nyohô-Akanagi, contamination by the argillaceous sediments as well as fractional crystallization were operative. The effect of contamination is seen most clearly in the enrichment of K_2O relative to Na_2O. This effect is distinguished from that of fractional crystallization by plotting the K_2O/Na_2O against SiO_2. The effect of contamination was more marked in the later stage of the volcano, resulting in the predominance of acidic andesite and dacite.

According to KUNO (1950), volcanic rocks having orthorhombic pyroxene in the groundmass (hypersthenic rock series) are formed from magmas contaminated by sialic materials, while those with pigeonite in the groundmass (pigeonitic rock series) are formed from uncontaminated magmas. The lavas of Series 3 of Nyohô-Akanagi,

which, from their chemical composition, is interpreted by the writer as having been formed from contaminated magmas, contain hypersthene in the groundmass and hornblende as phenocrysts. The lavas of Series 2 carry pigeonite in groundmass, yet their chemical compositions are indicative of their having been affected by contamination.

Since KUNO's definition of the two rock series is based on the presence of pigeonite or hypersthene in the groundmass, the lavas of Series 2 belong to the pigeonitic rock series (KUNO, 1950). Whether hypersthene or pigeonite is formed in the groundmass depends primarily on the relation between the temperature and the composition of the pyroxenic components of the magmas (KUNO, 1950). The present writer considers that the hypersthenic rock series is formed only when the magma is so markedly contaminated that physico-chemical condition becomes suitable for the formation of hypersthene. This view is not contradictory with KUNO's one, for KUNO does not seem to preclude the possibility of contamination during the formation of the pigeonitic rock series (KUNO, 1950 and 1953). It is possible then that the contamination in the lavas of Series 2 was not so marked as to form hypersthene in the groundmass.

The enrichment of K_2O relative to Na_2O in Series 2 and 3 may also be explained as due to lower degree of fractionation of plagioclase as compared with that in Series 1 (BOWEN, 1928 and KUNO, 1937). Which of the two alternative explanations suggested here are valid should be decided by further investigation.

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